ANNALS OF THE NEW YORK ACADEMY OF SCIENCES

VOLUME 95, ART. 1 PAGES 1–740

SOLAR VARIATIONS, CLIMATIC CHANGE, AND RELATED GEOPHYSICAL PROBLEMS

Conference Editor

RHODES W. FAIRBRIDGE

LIST OF AUTHORS

RHODES W. FAIRBRIDGE AND CHARLES G. KNUDSEN (*Conference Cochairmen*), R. Y. ANDERSON, G. BARTA, H. F. BECHER, A. BEISER, H. P. BERLAGE, F. BRANDTNER, G. W. BRIER, R. A. BRYSON, K. W. BUTZER, M. B. DAVIS, J. A. DUTTON, B. L. DZERDZEEVISKII, L. EGYED, C. EMILIANI, D. B. ERICSON, R. F. FLINT, H. FLOHN, H. FRIEDMAN, J. GENTILLI, E. D. GILL, J. B. GRIFFIN, G. HATTERSLEY-SMITH, C. J. HEUSSER, W. S. JARDETZKY, T. N. V. KARLSTROM, E. B. KRAUS, H. H. LAMB, D. B. LAWRENCE, E. G. LAWRENCE, A. LEROI-GOURHAN, G. MANLEY, J. M. MITCHELL, JR., G. N. PLASS, W. A. RENSE, P. B. SEARS, R. SHAPIRO, I. I. SCHELL, D. J. SCHOVE, R. SHUTLER, JR., T. L. SMILEY, R. S. SOLECKI, J. TERASMAE, TH. VAN DER HAMMEN, E. H. VESTINE, F. WARD, J. W. WARWICK, A. T. WATERMAN, H. C. WILLETT, D. WILLIAMS, E. H. WILLIS, L. W. WING, R. B. WOODBURY, H. E. WRIGHT, JR., AND F. E. ZEUNER

Editor

FRANKLIN N. FURNESS

Managing Editor

EDGAR W. WHITE

Associate Editor

MALCOLM EASTERLIN

NEW YORK

PUBLISHED BY THE ACADEMY

October 5, 1961

ANNALS OF THE NEW YORK ACADEMY OF SCIENCES

VOLUME 95, ART. 1 PAGES 1–740

October 5, 1961

SOLAR VARIATIONS, CLIMATIC CHANGE, AND RELATED GEOPHYSICAL PROBLEMS*

Conference Editor: RHODES W. FAIRBRIDGE

Conference Cochairmen: RHODES W. FAIRBRIDGE AND CHARLES G. KNUDSEN

Editor

FRANKLIN N. FURNESS

Managing Editor

EDGAR W. WHITE

Associate Editor

MALCOLM EASTERLIN

CONTENTS

* This series of papers is the result of a conference on *Solar Variations, Climatic Change, and Related Geophysical Problems* held and supported conjointly by The New York Academy of Sciences, New York, N.Y., and the American Meteorological Society, Boston, Mass., January 24, 25, 26, 27, and 28, 1961.

Part I. Astrophysics: Solar Activity and Variability

SOLAR INFLUENCES ON GEOMAGNETIC AND RELATED PHENOMENA

E. H. Vestine

Planetary Sciences Department, The Rand Corporation, Santa Monica, Calif.

Introduction

The geomagnetic field undergoes changes with time that are closely linked to time changes in solar phenomena. Bartels (1946) has shown that the correlation of sunspots with the annual average terrestrial equatorial changes in the geomagnetic field, indicated by the solar daily magnetic variation, is as high as 0.99. This is so high that it is believed that annual average values of certain fluctuations in solar-wave radiation are well monitored by measuring geomagnetic-field changes on the surface of the earth. The physical connection seems to be that the geomagnetic-field changes are due to electric currents flowing in the ionosphere; these currents are produced partly as a result of dynamo action of solar-driven upper-air winds. As is also well known, other aspects such as ion amount and temperature of the ionospheric regions are affected by solar-wave radiation (Chapman and Bartels, 1940).

It is found that certain transient geomagnetic changes and auroral displays in the polar regions seem linked to solar-corpuscular radiation. This is clear from inspection of K measures of magnetic disturbances of FIGURE 1 which, when more intense, depend on sunspot number. The physical connection is thought of as involving solar-proton streams propagated from sun to earth in about one day, but many of the actual features of such streams and their effects are obscure. These streams may already have been detected through the use of space probes. Thus preliminary findings on the Pioneer V solar probe have shown that magnetic fields appeared in space millions of miles from the earth; these fields later may have traveled to the earth, possibly being locked within a moving stream of solar protons. This suggests that solar streams of particles transported a magnetic field, and that these streams would in turn interact with the outer geomagnetic field (Coleman *et al.*, 1960). The effect of such interaction is not known, but compression of the geomagnetic field into a space 6 to 15 earth radii in the region of initial contact has been suggested (Chapman and Bartels, 1940; Warwick, 1959; Dessler and Parker, 1959; Matsuura and Nagata, 1960; and Beard, 1960).

Entry of Solar Particles into the Geomagnetic Field

It is known that high-energy solar protons of energies ranging up to those of cosmic rays readily penetrate the geomagnetic field, even to ground level in the case of cosmic rays. Other high-energy solar protons produce high absorption of radio waves in polar regions, due to the excess of absorbing electrons created within the low ionosphere (Bailey, 1957; Leinbach and Reid, 1959). It is also known from rocket measurements of electron impacts during auroral displays that low-energy particles such as 6-kev electrons are preponderant near the E region. These electrons are presumedly dumped from the Van Allen radiation

belts (Van Allen and Frank, 1959). They are thought to be generated indirectly by streams of solar protons, or by the effect of solar-proton streams upon plasma within the geomagnetic field (Dessler and Karplus, 1960). Some of these high-energy solar protons are of course trapped in the outer Van Allen radiation belts following solar flares but, in addition, it has been suggested that

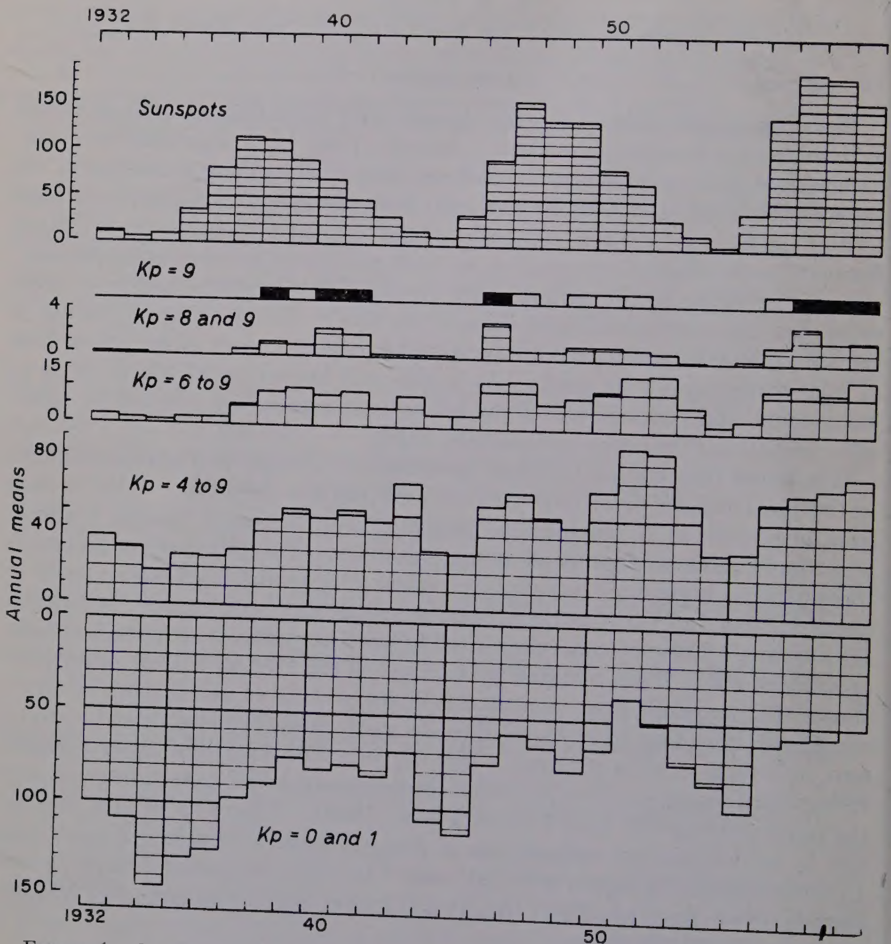


FIGURE 1. Sunspot numbers and frequencies of range indices, annual values. Adapted from Bartels (1946).

low-energy solar protons also diffuse into the Van Allen region from incident solar streams (Matsuura and Nagata, 1960; and others).

N. Matsuura and T. Nagata have recently estimated that unmagnetized solar streams may be able to penetrate the distorted geomagnetic field (Matsuura and Nagata, 1960). In order that diffusion occur, it is supposed that the boundary between the solar stream and geomagnetic field includes ionized clouds of average cross section about 1000 km., enduring for about 10 sec., enclosing

irregular magnetic fields. These authors then find that some of the protons and electrons will drift in the direction perpendicular to that of the crossed magnetic field and field gradient. In 10^4 sec., enough particles may enter the geomagnetic field to contribute to large geomagnetic-field changes known as magnetic storms; they endure for several days.

The cited authors did not consider a magnetized solar stream in their study. It may be noted, however, that such a magnetized solar stream may also permit particles to drift down to the Van Allen radiation belts by the same process. Rough estimates that I made suggest a similar efficiency of injection. Finally, as W. H. Bennett and E. O. Hulburt have suggested, there may be more concentrated segments of solar-proton streams directly entering the upper atmosphere mainly on the night side of the earth (Bennett, 1958). The Bennett–Hulburt beams apparently were not noted by Pioneer V, consequently either they do not exist continually or were missed by Pioneer V particle counters. It may also be remarked that the irregularly magnetized clouds in outer reaches of the Van Allen belts are only conjectural, although some magnetic-field roughness in that region has been noted by several space probes.

Consequently aurora and magnetic disturbances probably are a result of the interaction of solar streams with the geomagnetically trapped radiation of the Van Allen regions (Van Allen and Frank, 1959). This trapping may be rather transitory in character, and also long-continued, as in the case of the usual Van Allen radiation. In order that the aurora and polar electrojets be produced, some of the particles probably require increases in energy (Chamberlain *et al.*, 1960).

The way in which such radiation is dumped to form aurora and polar magnetic disturbances is not yet known. One possibility is that since the parts of the Van Allen radiation belts mirroring in high latitudes are not stable, fluted groupings of Van Allen radiation will form, given a sufficient length of time. Within these groupings transient electric fields that are transverse to the lines of geomagnetic force within the atmosphere should grow exponentially with the time, tending to drive segments of geomagnetically trapped charged particles poleward (Kern and Vestine, 1961). The tendency to instability may be greater on the night side where the lines of force may be stretched and the field weakened, in comparison with the day side of the earth, where compression of the geomagnetic field by the solar stream may be greater. It is possible that the flutes in the radiation belt that are created are further affected and shaped by magnetic and electric fields (Kern and Vestine, 1961; Vestine, 1960). If a pinch-type effect is possible, penetration to auroral levels may arise, with formation of aurora and electrojets of current flowing in the E region of the ionosphere. This is because of the diamagnetic action of any excess of electrons present, for example, in the radiation segment or flute, tend to lower the radiation mirror point slightly and, perhaps, facilitate the dumping of the radiation into the low ionosphere; in addition, a small component of any electric field present, if dumped along the geomagnetic field, would more likely be of importance in achieving significant dumping. In fact if the solar stream distorts the geomagnetic field on the night side, as suggested by Kern (1961) electrojets driven by Hall currents may form. At the present time attempts are being made to calculate such magnetic gradients; these, if directed away from the sun

on the dawn and afternoon sides of the earth, should afford separated charge aggregates that, if they can be dumped into the ionosphere, may provide an electric driving force for the polar electrojets. These electric forces will in turn tend to interact with the Van Allen radiation belts, and any longitudinal electric fields occasioned by compression or distension of the geomagnetic field will probably cause meridional drift of trapped Van Allen radiation. These various possible effects may be borne in mind while considering the various radiation shells. The latter will next be examined in more detail by use of a 48-term spherical harmonic expansion for the geomagnetic field. This material may be found useful in estimating the perturbations of the Van Allen belts caused by solar-proton streams. In this work, the magnetic field of particles moving in the radiation shells is neglected, but it will be considered in a later investigation. Also defined is the shell connecting the northern and southern average auroral zones. Near this shell, important transient effects in the radiation belts are noted. Some of these effects are attributed to the electric fields producing the polar electrojets and, therefore, ultimately to solar streams. The contribution to the injection of radiation into the outer levels of the Van Allen region, due to the electric fields of the polar electrojets, is suggested. An attempt is made to assess the importance of this contribution, using estimates of electric fields derived from known interactions of the electrojets with the F region of the ionosphere.

Surfaces of Particle Flow in Van Allen Regions

The existence of radiation in belts about a magnetized sphere was demonstrated in the laboratory by Birkeland (1908, 1913), but not explained and, no doubt, gave rise to the concept of a ring current encircling the earth at a distance of a few earth radii. The spiral motions and reflection of charged particles were computed mathematically by Störmer (1907). The drift of such spiraling particles in a magnetic-field gradient seems to have first been discussed by Gunn (1929). The modern development of these ideas with important extensions is due to Alfvén (1953) and, on the experimental side, to Bennett (1958) and others. The application of some of these ideas in the theory of magnetic storms was considered by Bennett, but in more detail by Singer (1957), who considered that low-energy particles might produce magnetic storms. The discovery that particles of even quite high energy existed in large numbers in trapped condition about the earth was soon thereafter made by Van Allen and his co-workers (Van Allen *et al.*, 1958). The Argus experiment proposed by Christofilos (1959) clarified and demonstrated these ideas by creating an artificial radiation shell by means of a small atomic explosion. It was also indicated that integral invariants of the particle motion should describe the surface within the geomagnetic field, defining the center of a region within which a charged particle is constrained to move (Northrup and Teller, 1960).

These surfaces are here shown in FIGURE 2, the surfaces being selected by the intersections of the orbit of the probe Pioneer III with surfaces on which the radiation counts were 10,000/sec. or 10/sec. (Van Allen and Frank, 1959). The surfaces of FIGURE 2 relate to motion, not to flux density, since the latter depends also upon the total magnetic field F (Ray, 1960; Northrup and Teller,

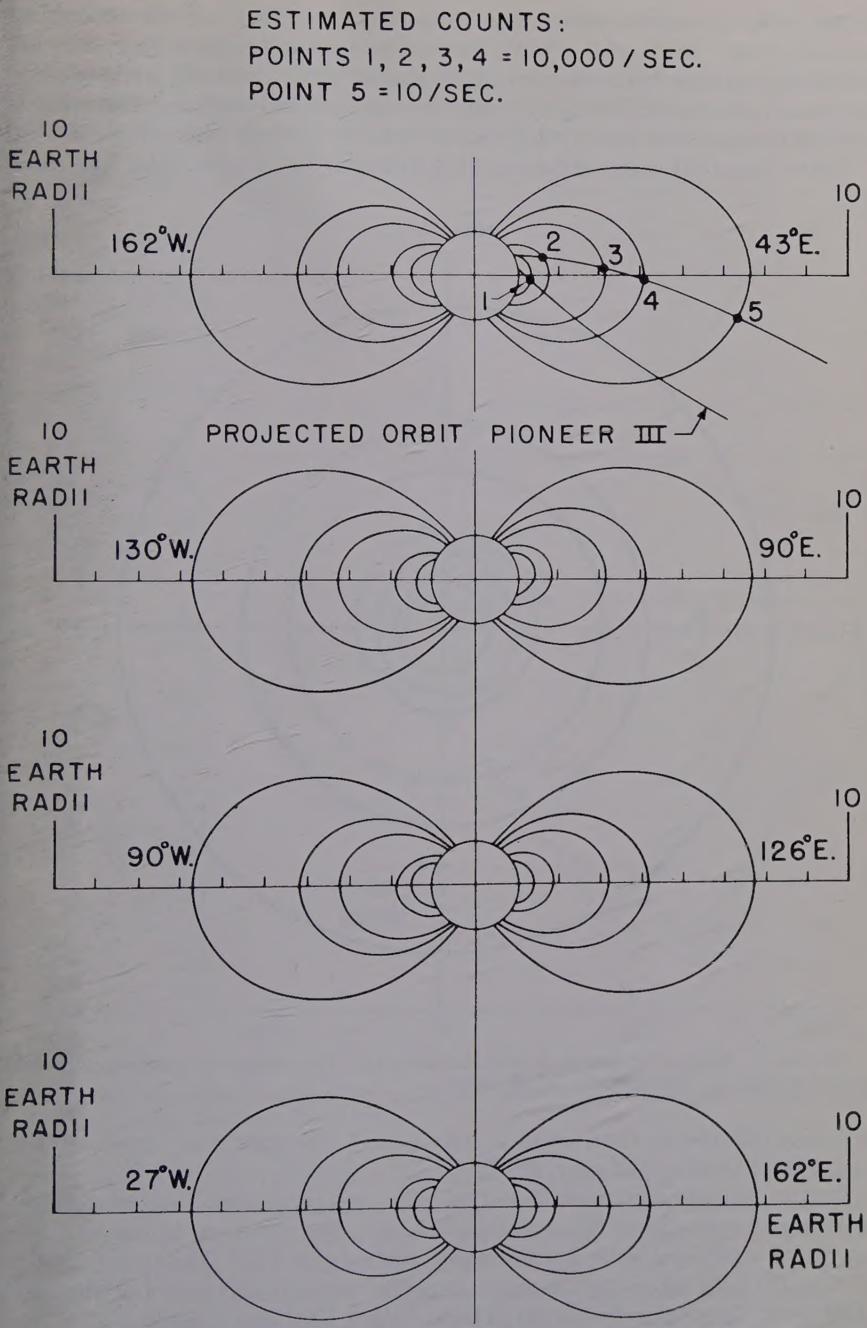


FIGURE 2. Computed projections of geomagnetic field lines on eight meridional planes.

1960). The outermost curves are intended to terminate at the theoretical auroral zones. Each shell (shown extending to ground level in FIGURE 3) actually may be regarded as terminating at a surface F = constant, a mirror point for those particles oscillating between the northern and southern hemispheres while drifting around the earth within or near the shell (Yoshida *et al.*, 1960).

Inspection of the upper surface passing through point 2 in the inner Van Allen

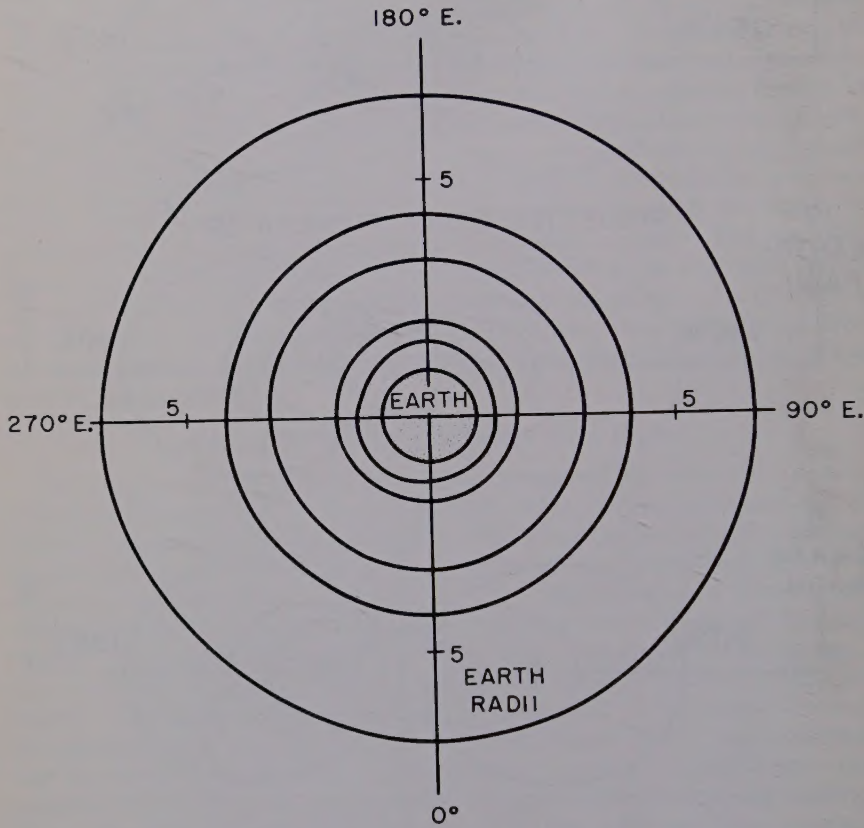


FIGURE 3. Equatorial section of drift shells for Van Allen radiation spiraling along field lines of FIGURE 2. Geomagnetic coordinates.

radiation belt reveals that, near lon. 160° W or E, the shell is at about 980 km. higher in elevation than near 40° E.

FIGURE 3 shows this result in another way, and the intersections of the shells of FIGURE 2 with the earth's equatorial plane. The results as before are based on the 48-coefficient spherical harmonic analysis for 1955 (Finch and Leaton, 1957). The actual maximum range in height was not obtained, but the inner shell in the figure was calculated to be at height 3300 km. above lat. 2° S, lon. 160° E, and 2318 km. above lat. 1° N, lon. 31° E, which gives an indication of the range. The heights of the mirror point F = 0.5 C. G. S. for the outer shell

are indicated in FIGURE 4, together with maximum heights and field values of the shell near the equator.

The geomagnetic annual and sunspot variations. FIGURE 5 shows the latitude distribution of the daily averages of geomagnetic disturbance for the 5 selected internationally disturbed days each month of 1922 to 1933 (Vestine *et al.*, 1947).

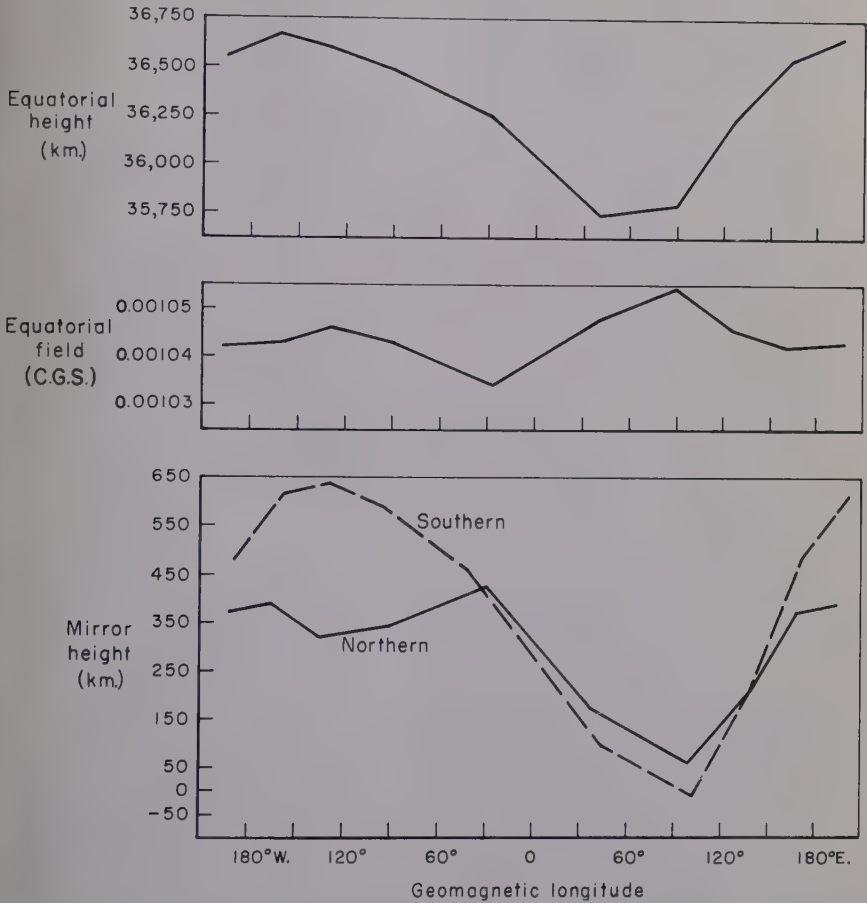


FIGURE 4. Maximum heights of lines of force at equator, field values; also corresponding high-latitude mirror heights of these lines.

The geomagnetic east component shows scarcely any variation with latitude between the northern and southern auroral zones. This is compatible with motion of particles along the surfaces or shells of FIGURE 3, approximately in the direction perpendicular to both the gradient of the geomagnetic field and the field itself. This adds credence to the view rapidly becoming established that averages of disturbances over periods of one day or more arise from drifting particles in the radiation belts.

Another feature is the added bulge in the geomagnetic north component ex-

tending for some distance to the north or south of the geomagnetic equator. This is of course likely due to the added average field of particles drifting in the high-energy (inner) Van Allen radiation belt.

The total current flowing from east to west could of course be easily estimated, for instance, by supposing the current flow in the outer and inner Van Allen radiation belts replaced by a thin shell near the maximum radiation density of each Van Allen region measured near the equatorial plane. If the ratios of the magnetic-field contributions by each shell are taken to be about 3 to 25, for example as seems reasonable from inspection of the data of FIGURE 5, one

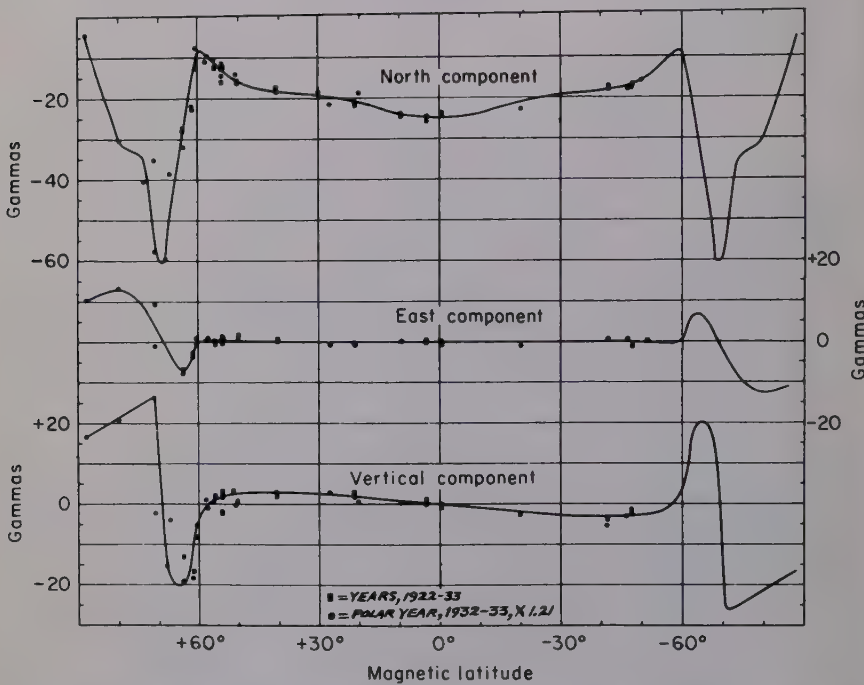


FIGURE 5. Variation with geomagnetic latitude in the daily averages of geomagnetic disturbance.

eighth of the equatorial part of the disturbance field averaged around parallels of latitude arises from the inner Van Allen radiation belt. This result seems to be in good qualitative agreement with Van Allen's results based on Pioneer III and IV, and should be checked when the average complete spectrum of energies becomes available.

The auroral-zone effects are known to be due to sources near the E region in the case of the polar electrojets, although a minor contribution to the electrojets may arise, as Dessler and Parker (1959) suggest, from the mirroring of particles from the radiation belt. These workers also point out that there is a contribution from the interhemispherical motion of the trapped particles, and they provide estimates of this contribution.

The Electric Field of the Polar Electrojet

In auroral regions, northern and southern, intensifications of the geomagnetic field of 500 to 1000 gammas (1 gamma = 10^{-5} C. G. S. U.) occur. These are called magnetic bays, and appear almost nightly, often in sequences of several nights, above the same locality, beginning at about the same hour (Chapman and Bartels, 1940). They last from about 1 to 5 hours. They are due to concentrated electrojets of current within or near the E region of the ionosphere at a height of about 100 km. The total current flowing in the electrojet is estimated to be about 500,000 amp., but may rise to values in excess of 1,000,000 amp. (Vestine *et al.*, 1947). The actual current cross section has not yet been measured, but may be several hundred kilometers wide and perhaps 50 km. thick; the distribution of current might also be that for auroral arcs. The electric conductivity of the region is normally augmented by aurora. If the conductivity is taken to be about $10^2 \times$ normal, electric potentials of less than one tenth mv/cm. directed from north to south should drive the early morning electrojet. This potential arises from dynamic effects associated with the aurora and, perhaps, also from upper-air winds.

Obayashi (1959) has recently summarized results of his own and other studies of the influence of the polar electrojets and other features of magnetic storms upon the ionosphere. Although certain features of the interrelationship between the phenomena are obscure, it seems likely that the electric field of the polar electrojets raises or lowers the F region of the ionosphere (Martyn, 1953).

The effect of the electrojet on the F region varies with the intensity of magnetic disturbance, and may be great enough to blow the F region from the upper atmosphere within one hour (Berkner and Seaton, 1940). Ordinarily there are only modest changes in height with a rise in the F region of some tens of kilometers. A rise in height, or lowering of the critical frequency f_0F_2, is common on the forenoon side of the earth and may be accompanied by a lesser transient depression in height at night.

The F region may rise or fall in response to the motor effect of electric currents. The F region may also move in response to a crossed electric and magnetic field. If its velocity is v, this may be regarded as the driven velocity of ions and electrons (Martyn, 1953). If for the moment only the electrons are regarded, the current i is given by

$$i = Nev = \sigma_2 h \times E/H \tag{1}$$

where i is the current, N the charge density of electrons (neglecting positive charges for the moment, since we are interested mainly in motion of electrons only), e the electronic charge, and v the velocity across the field lines. Also σ_2 is the electric Hall conductivity, h a unit vector in the direction of the magnetic field H, and E the electric field. Since

$$\sigma_2 = \frac{Ne^2}{m\nu} \frac{\nu w}{\nu^2 + w^2} \tag{2}$$

where ν = collisional frequency and w the spiral frequency, He/m. When $w > \nu$, as is true is the upper ionosphere and above, then very nearly

$$\sigma_2 \sim NeE/H \tag{3}$$

and from (1) $Nev = NeE/H$ or

$$v = E/H \sim 3ER^3 \tag{4}$$

since $H \sim 0.3/R^3$ in low latitudes, and is very nearly the same for ions.

Because of the decrease in H with height, v increases rapidly as R^3, while E is reduced with height because of electromagnetic induction. If $H = 0.5$ and $vt = 40$ km. where t is 4000 sec., $E = Hv = 0.5 \times 40 \times 10^5/2000 = 10^3 =$ emu $= 10^{-5}$ v/cm. Above R is distance from the earth's center to the F region measured in earth radii. Of course, even if a very feeble current flows across the field, a noticeable motor effect may occur and even dominate in determining the motion of the electrons, whether ambient as in the F region or in the form of Van Allen radiation within the exosphere. If the upward penetration of the electrojet fields, including those closing the electrojet circuits, were calculated which appears, unfortunately, to be a matter of considerable difficulty, more precise estimates could be made of the charges in the radiation belt caused by the polar electrojets. In this connection it may be noted that electrojets may closely simulate an ideal electric doublet, or appear as the superposition of many such doublets distributed along the auroral zone. From data on transmission of geomagnetic fluctuations in the ionosphere, theoretically estimated by J. H. Piddington, it is clear that electrojets of short duration, for example one hour or less, cannot provide an electromagnetic signal penetrating very far into the ionosphere; consequently the outer Van Allen region should not be affected (Piddington, 1959). It may also be that additional electric fields are present, as Alfvén (1953) has supposed. A sharply defined electrojet may also, under certain conditions, be associated spatially with auroral rays, within which at least a part of the electrojet actually flows. The electrojet and auroral display may then move south together (Heppner, 1954). If such a jet is formed of down-coming solar particles, some of which are reflected, the latter may be able to enter a trapped condition in the outer Van Allen radiation belt. Since on relatively quiet days the auroral zone is farther poleward, a greater radial extent of the outer Van Allen region during quiet periods might be expected. There appear to be some reasons for believing that this may be the case (Van Allen *et al.*, 1959; Jastrow, 1959; and Vestine, 1960).

Other Effects of Electric Fields Originating in the Polar Ionosphere

It has been noted that with the aid of a few satellite radiation measurements and calculations of the geomagnetic field above the ground, the approximate geometry of various radiation shells at any given time can be inferred, using the theory of drifting trapped radiation. This expectation is justified for the inner Van Allen belt, within which the geomagnetic field is of adequate intensity to maintain high stability. This may also be true for much of the outer belt, but there is a much greater chance of disturbing effects, such as those due to solar-stream fluctuations because the geomagnetic field is less intense at high equatorial elevations. The latter disturbing effects arise also from the propagation of electromagnetic effects upward from the low ionosphere, or downward from the hypothetical boundary between the solar stream and the geomagnetic field.

It is known from geomagnetism that transverse electric fields occur within

the ionosphere. These electric fields may grow exponentially with time at a rate such that an e-fold increase occurs in the time

$$t = \sigma_1 AH/4Lp\nu_d \quad (5)$$

where σ_1 = the transverse electric conductivity, A the ionospheric electrojet-current cross section, H the magnetic field, L the length of field lines between mirror points, p the charge density, and ν_d the drift velocity (Kern and Vestine, 1961). An atmospheric wind-produced irregularity in electric field, for example, may possibly also initiate flutes within the terminating region of a given radiation shell. The effect is mainly of shift of a segment of the base of the shell, as a flute, along its entire interhemispherical length into an adjacent shell. This may be one possible cause of the remarkable number of irregularities in radiation counts noted by earth satellites within the outer radiation belt. At night, when the transverse conductivity σ_1 is less, the time constant for growth of flutes is shorter. Some of these problems have been otherwise treated recently by others (Dungey, 1959; Gold, 1959; and Parker, 1960).

It may be remarked that longitudinal gradients in the geomagnetic field near the boundary between the solar stream and the geomagnetic field and due to fluctuations in the compressive and distorting influence of the solar streams will cause the shells of electrons or protons in the radiation belts to drift in a direction perpendicular to both the geomagnetic field and the magnetic-field gradient. This drift is opposite for protons and electrons. It seems likely an important reason for migration of charge aggregations in the outer Van Allen radiation belt, or even into the gap between the outer and inner belts. Within the latter region, differences in charge distribution within the outer and inner radiation belts may arise from longitudinal magnetic-field gradients (Kern, 1961). It would be of high interest to know if there may arise a resultant component of electric field directed along the geomagnetic field that would cause dumping of trapped or other particles. This mechanism may also contribute to the gap between outer and inner belts; qualitatively speaking, the protons would respond to the east–west longitudinal gradient by drifting outward from the earth in the outer-radiation belt, and some electrons would drift downward above the equator.

Oscillations in the direction of the longitudinal magnetic gradient in the equatorial plane would of course give rise to oscillatory effects often noted in the polar electrojets flowing in auroral regions of the ionosphere. If these theories are correct, a knowledge of the polar-disturbance field should permit estimates of the distortion of the geomagnetic field in nearby space from ground-based observations. There is need for a quantitative discussion in order to assess the importance of this effect, as well as of the electric acceleration responsible for the dumping of auroral particles; the latter will also depend upon and affect the spectrum of energies present and the numbers of particles in trapped condition.

Summary

Several shells, along which a particle trapped in the Van Allen radiation belts must move, are defined by means of calculations of the geomagnetic

field and integral invariants of the particle motion for 1955.0, neglecting the influence of ring-current magnetic fields. Such a shell passing through the inner Van Allen radiation belt is found to be at vertical height 2320 km., at lat. 1° N, lon. 31° E. The height is 3300 km. at lat. 2° S, lon. 160° E. The height of the 0.5-gauss mirror points for particles approaching the auroral zones in various longitudes is indicated.

The polar electrojets may arise from distortions of the geomagnetic field by solar streams, providing a dumping of auroral particles trapped temporarily and accelerated within outer reaches of the Van Allen radiation belt or region; longitudinal magnetic field gradients should separate proton and electron shell segments. The strong nighttime electrojet should then be driven by the approximately north–south electric field of the dumped particles; the electrojet of the early evening should arise from the south–north electric field. These polar electrojets, which are located at the northern and southern auroral zones, probably in turn distort adjacent portions of the radiation belts because of the electric fields of the electrojets, but it is difficult to estimate the importance of this effect. During magnetic storms, the penetration of the electric field driving the electrojets extends as a return circuit to higher levels above the earth, since the field change is of longer duration. For the morning electrojet, the electrostatic field closing the ionospheric current circuit is likely to be from west to east within lower reaches of the belts, and should drive the radiation upward and poleward, so that some of it may be lost into space. The electrojets are farther equatorward during a storm, so that a wide section of the upper part of the outer radiation belt might be removed. A more important source of electric field may arise from transient compression and rarefaction of the geomagnetic field by solar streams. This should provide a longitudinal component of electric field in the outer Van Allen region causing poleward or equatorward drift of radiation shells.

During quieter periods, the electrojets, now farther poleward, endure for so short a time (one hour or so) that penetration of their electric fields is insignificant at radiation-belt level. The general effect may be therefore summarized by remarking that in the case of minor magnetic disturbances the distribution of electrical driving forces will be mainly a two-dimensional skin effect confined to the lower ionosphere, whereas in the case of large enduring disturbances there may be deep though feeble penetration into the Van Allen regions as well.

Longitudinal geomagnetic-field gradients due to streams of solar protons will be likely to cause local distortions in the radiation belts. The transfer of charge from magnetized solar streams of 50-kev energy to the geomagnetic field seems most likely just after dawn, in the case of protons, and in the late evening for electrons, on the basis of calculations by Matsuura and Nagata (1960); in this case the particle radiation in the belts will then undergo an additional drift as a whole to the east.

Quite near the electrojets, the gradients in the combined main and electrojet magnetic fields will cause complex drifts to occur in the case of downcoming beams of auroral particles. These fields may have pronounced effects in shaping auroral forms caused by the downgoing or upgoing particles spiraling in the geomagnetic field.

It is suggested that about one eighth of the current responsible for the annual and sunspot variations in geomagnetism flows in the inner Van Allen radiation belt.

Acknowledgment

I gratefully acknowledge the skilled assistance of W. L. Sibley, who did the machine analysis of the mathematics of this paper.

References

Alfvén, H. 1953. Cosmical Electrodynamics. Oxford Univ. Press. London, England.
Bailey, D. K. 1957. Disturbances in the lower ionosphere observed at VHF following the solar flare of 23 February 1956 with particular reference to auroral-zone absorption. J. Geophys. Research. **62:** 431–463.
Bartels, J. 1946. Geomagnetic data on variations of solar radiation: Part 1—wave-radiation. Terrestrial Magnetism and Atmospheric Elec. **51:** 181–242.
Beard, D. B. 1960. The interaction of the terrestrial magnetic field with the solar corpuscular radiation. J. Geophys. Research. **65:** 3559–3568.
Bennett, W. H. 1958. Auroral and magnetic-storm theory. Astrophys. J. **127:** 731–742.
Berkner, L. V. & S. L. Seaton. 1940. Ionospheric changes associated with magnetic storm of March 24, 1940. Terrestrial Magnetism and Atmospheric Elec. **45:** 393–418.
Birkeland, K. 1908–1913. On the cause of magnetic storms and the origin of terrestrial magnetism. *In* Norwegian Aurora Polaris Expedition, 1902–1903. **1** (Part 1): 39–315; (Part 2): 319–551.
Chamberlain, J. W., J. Kern & E. H. Vestine. 1960. Some consequences of local acceleration of auroral primaries. J. Geophys. Research (Letters to the Editor). **65:** 2535–2537.
Chapman, S. & J. Bartels. 1940. Geomagnetism. 2 vols. Clarendon Press. Oxford, England.
Christofilos, N. C. 1959. The Argus experiment. J. Geophys. Research. **64:** 869–875.
Coleman, P. J., Jr., C. P. Sonett, D. L. Judge & E. J. Smith. 1960. Some preliminary results of the Pioneer V magnetometer experiment. J. Geophys. Research. **65:** 1856–1857.
Dessler, A. J. & R. Karplus. 1960. Some properties of the Van Allen radiation. Phys. Rev. Letters. **4:** 271–274.
Dessler, A. J. & E. N. Parker. 1959. Hydromagnetic theory of geomagnetic storms. J. Geophys. Research. **64:** 2239–2252.
Dungey, J. W. 1959. The effect of electric fields on Van Allen's trapped radiation. Symposium on Physical Processes in the Sun-Earth Environment. Radio Physics Laboratory, Ottawa. July 20–21. Unpublished.
Finch, H. F. & B. R. Leaton. 1957. The earth's main magnetic field—epoch 1955.0. Monthly Notices Roy. Astron. Soc. **7:** 314–317.
Gold, T. 1959. Motions in the magnetosphere of the earth. J. Geophys. Research. **64:** 1219–1224.
Gunn, R. 1929. Preliminary note on the origin of the permanent magnetic fields of the sun and earth. Terrestrial Magnetism and Atmospheric Elec. **34:** 154.
Heppner, J. P. 1954. Time sequences and spatial relations in auroral activity during magnetic bays at College, Alaska. J. Geophys. Research. **59:** 329–338.
Jastrow, R. 1959. Density and temperature of the upper atmosphere. Astronautics. **4:** 24.
Kern, J. W. 1961. Solar stream distortion of the geomagnetic field and polar electrojets. J. Geophys. Research. **66:** 1290–1292.
Kern, J. W. & E. H. Vestine. 1961. Theory of auroral morphology. J. Geophys. Research. **66:** 713–723.
Leinbach, H. & G. C. Reid. 1959. Ionization of upper atmosphere by low-energy charged particles from a solar flare. Phys. Rev. Letters. **2:** 61–63.
Martyn, D. F. 1953. Electric currents in the ionosphere. III. Ionization drift due to winds and electric fields. Phil. Trans. Roy. Soc. London. Ser. **A246:** 306–320.
Matsuura, N. & T. Nagata. 1960. Paper III. On the earth storms. Interaction between the solar corpuscular stream and the earth's magnetic field. Report of Ionosphere and Space Research in Japan. **14:** 259–272.
Northrop, T. G. & E. Teller. 1960. Stability of the adiabatic motion of charged particles in the earth's field. Phys. Rev. **117:** 215–225.

Obayashi, T. 1959. Geomagnetic storms and ionospheric disturbances. J. Radio Research Labs. Japan. **6:** 375–514.

Parker, E. N. 1960. Geomagnetic fluctuations and the form of the outer zone of the Van Allen radiation belt. J. Geophys. Research. **65:** 3117–3130.

Piddington, J. H. 1959. The transmission of geomagnetic disturbances through the atmosphere and interplanetary space. Geophys. J. Roy. Astron. Soc. **2:** 173–189.

Ray, E. C. 1960. On the theory of protons trapped in the earth's magnetic field. J. Geophys. Research. **65:** 1125–1134.

Singer, S. F. 1957. A new model of magnetic storms and aurorae. Trans. Am. Geophys. Union. **38:** 175–190.

Störmer, C. 1907. Sur les trajectoires des corpuscules électrisés dans l'espace sous l'action du magnétisme terrestre avec application aux aurores boréales. Arch. Sci. Phys. et Nat., Bureau des Archives, Gèneve.

Van Allen, J. A. & L. Frank. 1959. Radiation around the earth to a radial distance of 107,400 km. Nature. **183:** 430–434.

Van Allen, J. A., G. H. Ludwig, E. C. Ray & C. E. McIlwain. 1958. Observation of high intensity radiation by satellites 1958 alpha and gamma. Jet Propulsion. **28:** 588–592.

Van Allen, J. A., C. E. McIlwain & G. H. Ludwig. 1959. Radiation observations with satellite 1958 epsilon. J. Geophys. Research. **64:** 271–286.

Vestine, E. H. 1960. Polar auroral, geomagnetic, and ionospheric disturbances. J. Geophys. Research (Letters to the Editor). **65:** 360–362.

Vestine, E. H., I. Lange, L. LaPorte & W. E. Scott. 1947. The Geomagnetic Field, Its Description and Analysis. Carnegie Inst. Wash. Publ. 580. Washington, D. C.

Warwick, J. W. 1959. Some remarks on the interaction of solar plasma and the geomagnetic field. J. Geophys. Research. **64:** 389–396.

Yoshida, S., G. H. Ludwig & J. A. Van Allen. 1960. Distribution of trapped radiation in the geomagnetic field. J. Geophys. Research. **65:** 807–813.

HYDROMAGNETIC WAVES IN THE UPPER ATMOSPHERE AND SOLAR ACTIVITY

Arthur Beiser
New York University, New York, N.Y.

In general, disturbances can propagate in a hydromagnetic medium in three modes. The best known of these is the so-called Alfven, or transverse mode, which is a noncompressive disturbance traveling in the direction of the field line. It may be understood intuitively by comparing the field lines to strings under tension. Here the inertia corresponding to the mass of the string comes from the fact that in a highly conducting gas the material and the field lines move together. Thus a transverse wave in a hydromagnetic medium is like a pulse propagating along a string. The other two modes, commonly called fast and slow, are combinations of gas dynamical effects and effects due purely to the existence of the magnetic field. If we suppose a compressive wave traveling in a direction normal to that of the field, since the field and the material move together the field lines will be compressed where the material is compressed. The compression of the field lines leads to an increase in the potential energy involved in the compression and so to an increase in the velocity of the wave over that of a pure sound wave. In the upper atmosphere, the velocity of an Alfvén wave is so much greater than the velocity of a sound wave that the wave velocity for the fast mode is the same in all directions. The slow mode, on the other hand, propagates largely along the field lines with some three dimensional spreading. Alfven waves, then, travel along the lines of force of the magnetic field with a characteristic speed depending upon the ambient ion density and the magnetic flux density; fast waves may travel in any direction relative to the field and have a speed greater than the speed of sound but less than the Alfven speed, though in the outer atmosphere the fast and Alfven speeds are essentially identical; and slow waves, which also may travel in any direction but whose speed is less than the sound speed. FIGURE 1 is a plot of V/V_s versus $S = V_A/V_s$, where V is the wave speed, V_A the Alfven speed, and V_s the sound speed.[1] In a medium of finite conductivity, the damping of the fast mode is independent of direction of propagation and is the same as that of the Alfven mode which travels along the field lines. The slow mode, however, is rapidly damped when its direction of propagation deviates from that of the field lines.

In a recent paper Francis and Karplus[2] have studied the absorption of hydromagnetic waves vertically incident on the atmosphere from an altitude of 550 km. Their treatment was restricted to a magnetic dip angle of 60° corresponding to a magnetic latitude of about 45°. They also invoked ionospheric properties corresponding to the daytime at sunspot maximum. An incident-hydromagnetic wave separates into an ordinary wave whose properties are independent of the direction of the magnetic field and an extraordinary wave. These two modes are coupled by the Hall effect, which is most important near an altitude of 120 km. FIGURE 2 shows the power dissipation versus altitude above the earth's surface for 2 ordinary waves whose frequencies are 25 radians/

sec. and 1 radian/sec. at normal incidence with their amplitude initially 100 γ. Evidently the power dissipation occurs in a relatively narrow altitude band. FIGURE 3 shows power dissipation versus altitude for extraordinary waves of 25 and 1 radian/sec. incident frequency, also with a 100 γ field amplitude initially. The ionosphere is essentially transparent for frequencies less than 1 radian/sec.

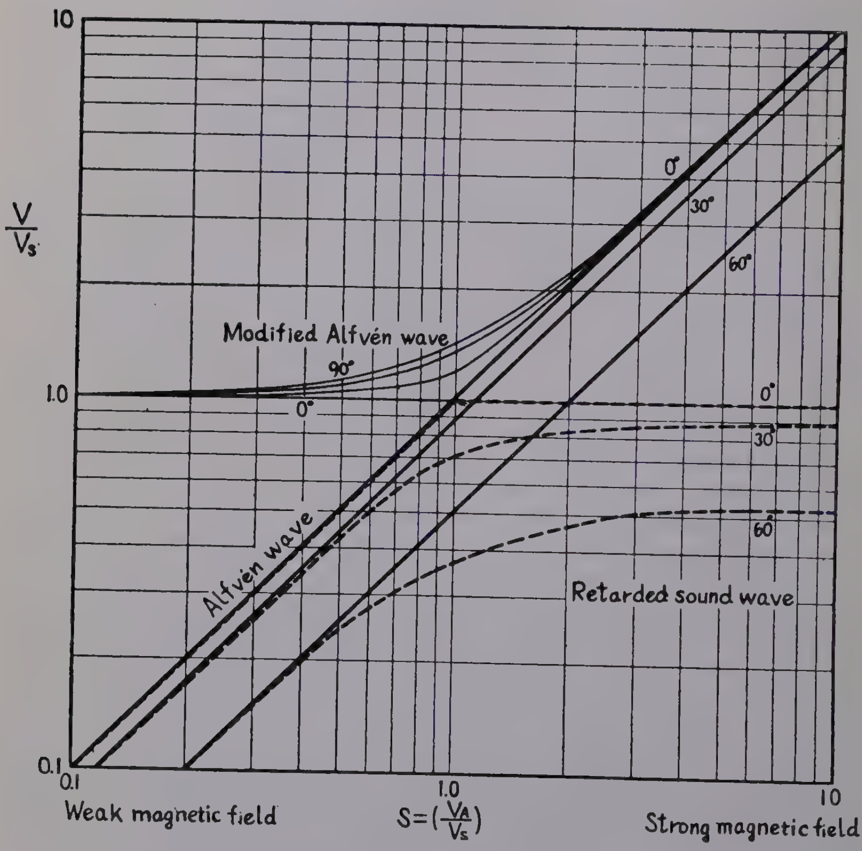


FIGURE 1.

Dessler[3] has pointed out that the energy dissipated in the ionosphere by hydromagnetic waves may represent an important source of ionospheric heating. This effect has been invoked by various authors to explain the following phenomena: (1) the lifting of the F region during geomagnetic storms, (2) the irregular orbital acceleration of satellites, (3) the sudden disappearance of trapped radiation from the Argus nuclear explosion coincident with a geomagnetic storm, (4) the X-ray flux observed at balloon altitudes below the Van Allen belt during geomagnetic storms, and (5) the decrease in intensity at the lower edge of the Van Allen radiation belt during geomagnetic storms.

It is interesting to note that the curves showing power absorbed versus alti-

tude for solar ultraviolet radiation and hydromagnetic waves are remarkably similar above an altitude of about 100 km.

We now turn to possible sources of hydromagnetic waves that may be incident upon the earth.

The sun constantly emits a totally ionized hydrogen plasma, usually called the *solar wind*, whose flux at the earth's orbit is perhaps 10^{13} ions/m.^{2}-sec. The kinetic energy of the protons is about 10^{-15} joule/proton, and the resulting energy flux at the earth's orbit is therefore in the vicinity of 10^{-2} watts/m.^{2}, about 10^{-5} of the flux of solar radiant energy.

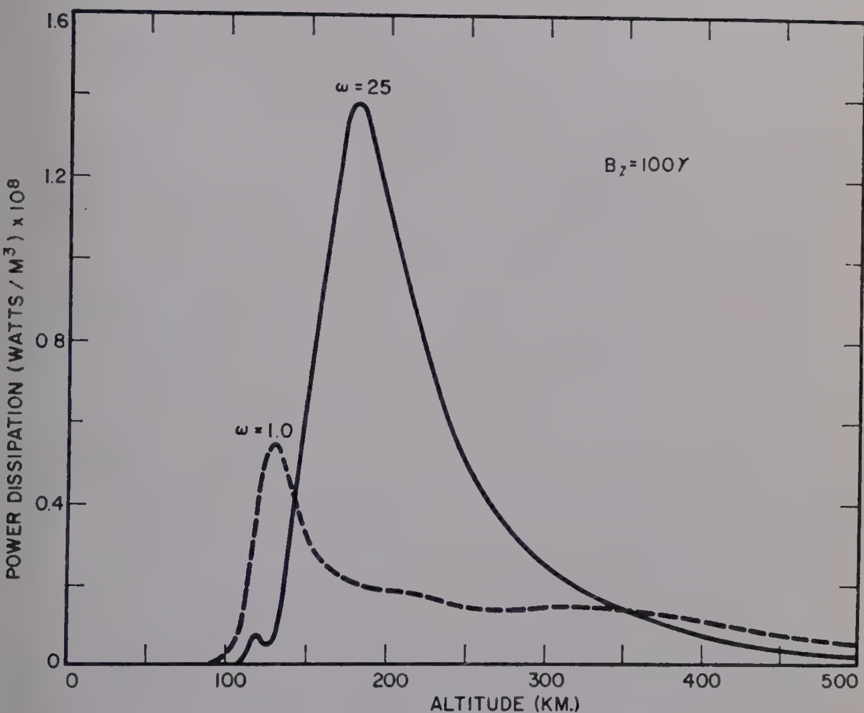


FIGURE 2.

The solar wind impinging on the earth's magnetic field causes it to be confined within a cavity in the wind. Estimates of the energy density of the plasma vary, but it seems likely that the plasma and field energy densities are equal at between 6 and 9 earth radii. The problem in three dimensions is too complex for exact solution. Instead, we have made an exact calculation of the cavity shape in 2 dimensions, corresponding to a pair of antiparallel line currents immersed in a plasma stream.[4] The result in the case where the dipole axis is perpendicular to the stream direction is shown in FIGURE 4; note that the neutral points lie somewhat upstream from the dipole axis. The maximum diameter of the cavity at infinity is twice that at the axis. Also shown is the cavity surface as calculated on the basis of an approximate pro-

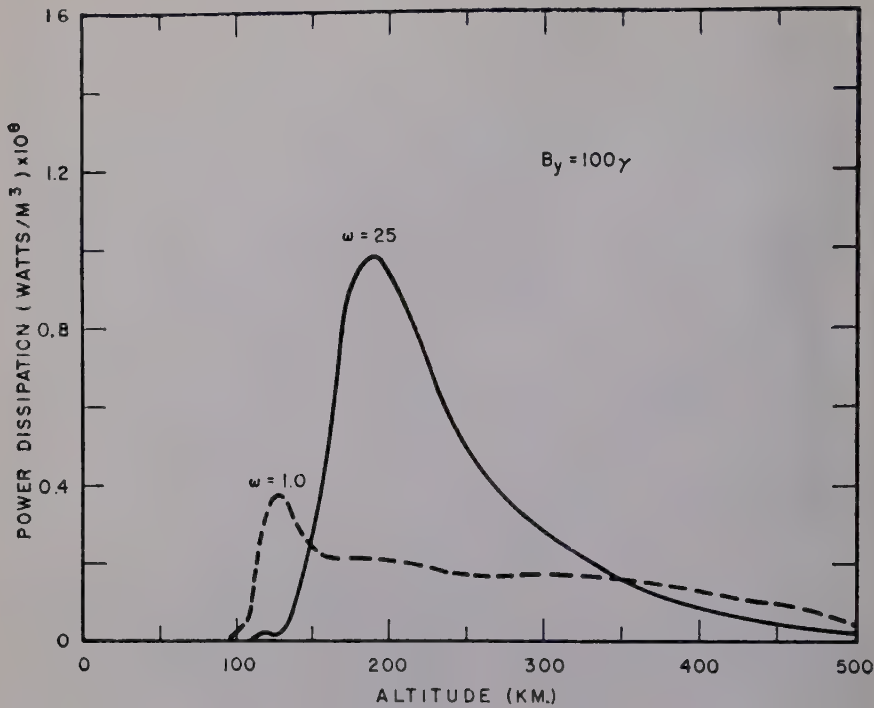


Figure 3.

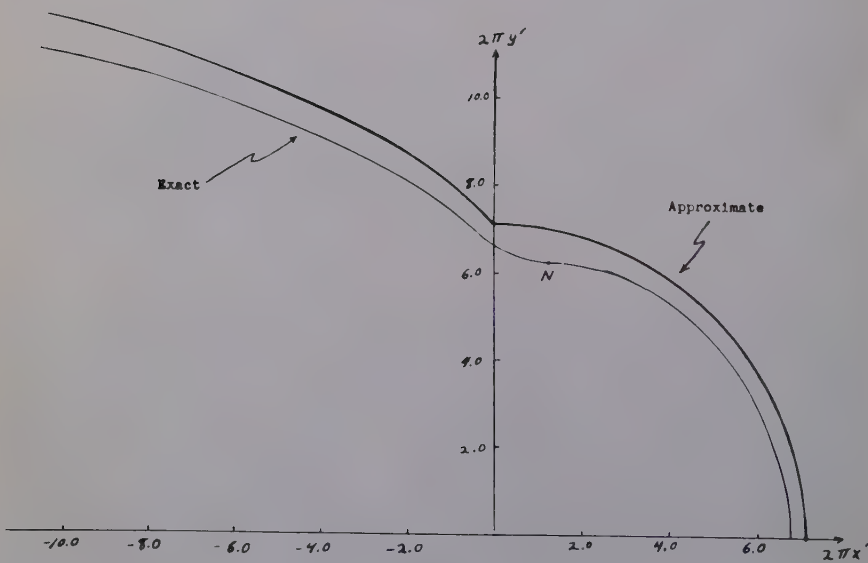


Figure 4.

cedure devised by Beard for the three-dimensional case: the agreement is sufficiently good for it to be assumed that Beard's method is probably satisfactory for the 3-dimensional problem. FIGURE 5 shows how the cavity is distorted when the dipole axis is tilted toward the stream by 37°, corresponding to the maximum tilt of the geomagnetic axis toward the sun. FIGURE 6 shows the result of an exact calculation of the field configuration within the cavity. (Note that we have not included the thermal effects within the solar wind that will cause the cavity to pinch down at infinity.)

Taking the radius of the cavity nose to be 10 earth radii, its cross-sectional

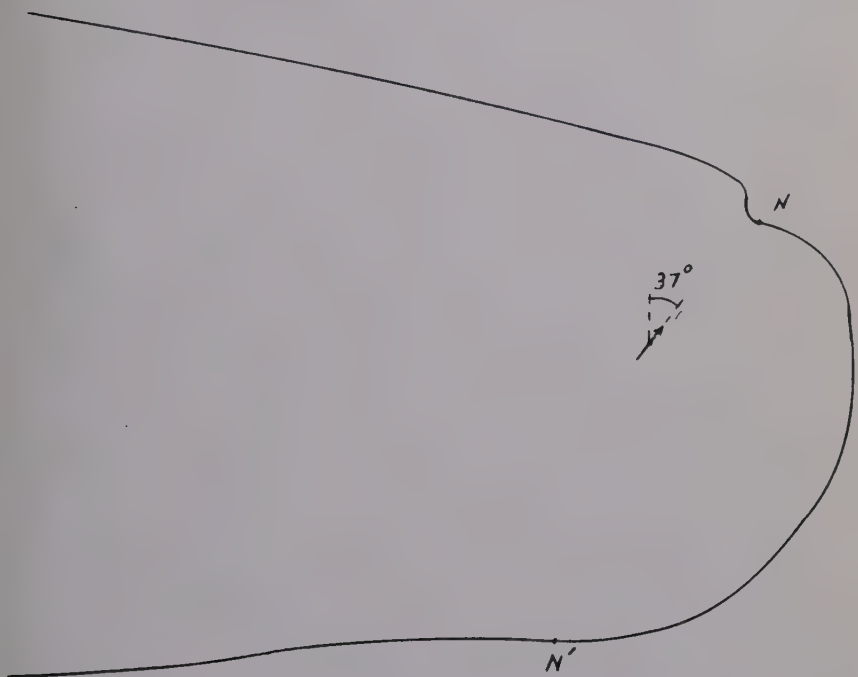


FIGURE 5.

area is 10^{16} sq. m. Thus the total power intercepted by the cavity nose is about 10^{14} w. By comparison, the total solar electromagnetic radiation striking the earth is about 2×10^{17} w. The cross-sectional area of the cavity downstream is 4 times greater than that of the cavity nose, so it intercepts 4 times more power. Even so there is not much total energy influx, but its selective absorption may make it more significant than these numbers indicate.

Hydromagnetic waves can be initiated at the cavity wall by two distinct processes: fluctuations in the incident solar wind and instabilities in the cavity wall. In the first case, the impact of irregularities in the solar wind on the geomagnetic field will cause the generation of hydromagnetic waves over the nose of the cavity surface. The result will be large amplitude hydromagnetic waves starting on the cavity surface and propagating inward. The arrival of

these waves at the earth's surface has been thought[5] to result in worldwide sudden commencement, a notion that has been claimed to give reasonably good agreement with the observed features of sudden commencements (other evidence[6] casts some doubt on this interpretation). The hydromagnetic waves that leave various portions of the surface will arrive at a given point on the earth at different times owing to their different paths. Furthermore, the impact of an initially plane plasma front on the field does not occur simultane-

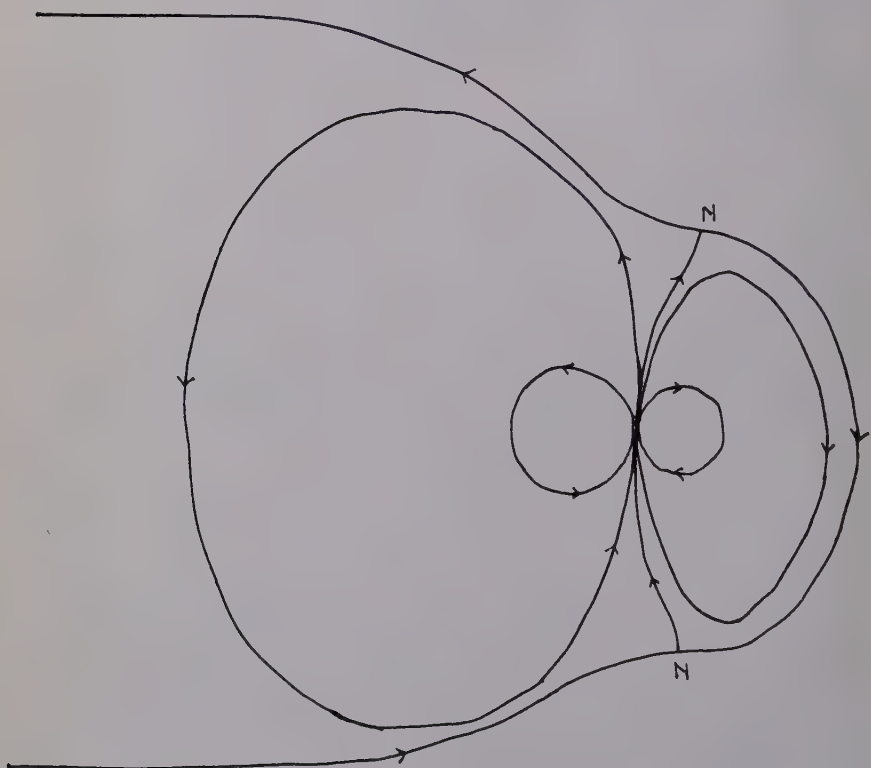


FIGURE 6.

ously along the curved geomagnetic surface but occurs first near the sun-earth line. Thus hydromagnetic waves are not generated simultaneously along the nose. These two effects produce a spread in the arrival times, at an observing point on the earth, of signals generated over the entire nose. The resulting delays yield a build-up of the sudden commencement amplitude with a rise time of roughly one to six minutes.

The second source of hydromagnetic waves is, in a way, the most interesting for our purposes. As a general rule, the cavity wall is stable where its center of curvature lies inside the cavity. Hence the nose is stable; but the regions near the neutral points are not, and the entire downstream part of the wall is

not. We are currently studying the propagation of hydromagnetic waves propagating inward from these regions.

In this connection it is interesting to speculate on the possibility of a duct within the upper atmosphere within which hydromagnetic waves may propagate unimpeded around the earth. The velocity of hydromagnetic waves in the atmosphere starts at about 227 km./sec. at an altitude of 200 km., drops to a minimum of perhaps 137 km./sec. at an altitude of 400 km., then increases once more to a maximum of 5,180 km./sec at an altitude of 3,000 km., and then declines steadily with increasing altitude. Thus refraction effects may cause a wave guide to exist for hydromagnetic waves at an altitude of several hundred km. above the earth. Detailed calculations are in progress on this point.

The work referred to above that has been conducted at New York University received support from the National Science Foundation, the National Aeronautics and Space Administration, Washington, D. C., and the Advanced Research Projects Agency of the Department of Defense, Washington, D. C.

References

1. Obayashi, T. 1958. Rept. Ionosphere Research Japan. **12:** 301.
2. Francis, W. E. & R. Karplus. 1960. J. Geophys. Research. **65:** 3593.
3. Dessler, A. J. 1959. J. Geophys. Research. **64:** 397.
4. Hurley, J. 1961. Doctoral thesis. Department of Physics. New York Univ. New York, N.Y.
5. Dessler, A. J., W. E. Francis & E. N. Parker. 1960. J. Geophys. Research. **65:** 2715.
6. Troitskaya, V. A. 1961. J. Geophys. Research. **66:** 5.

SOLAR VARIABILITY IN X-RAY AND ULTRAVIOLET EMISSIONS OBSERVED BY MEANS OF ROCKETS

Herbert Friedman
United States Naval Research Laboratory, Washington, D.C.

Introduction

In their efforts to understand solar terrestrial relationships, geophysicists and solar physicists are constantly seeking for indices of solar activity that can be related to variable phenomena in the terrestrial atmosphere. The indices take the form of: sunspot numbers; plage areas; coronal green, red, and yellow line intensities and their distributions over the disk; fluxes of solar radio noise; and the brightness, area, and line widths of solar flares. Such indices give clues from which one may infer the mechanisms of solar activity phenomena and the extreme ultraviolet and X radiations that accompany them, but it is impossible to be certain of any theory of solar activity without directly determining the ultraviolet and X-ray spectra. Scattered rocket measurements of these radiations have been accumulating over approximately one solar cycle and, within the past year, the first satellite observatory for monitoring X-ray and Lyman-α emission was successfully orbited. These experiments reveal a great variability in solar X-ray emission. The information regarding ultraviolet activity is not conclusive, but there is indirect evidence for large variations.

X-Ray Emission from the Quiet Sun

The sun's X-ray emission originates in the corona and perhaps in certain active regions of the chromosphere. The temperature of the corona is indicated by a variety of characteristic features: (1) the gravitation equilibrium; (2) the absence of Fraunhofer lines in the white light corona; (3) the intensities of the coronal emission lines arising from highly stripped ions of the heavy atoms; (4) the Doppler broadening of the coronal lines; (5) the absence of low-temperature spectral lines typical of the chromosphere and prominences; and (6) the radio noise spectrum. All of these phenomena are indicative of coronal temperatures in the range from about 5×10^5 to 2×10^6 °K.

Since the only X-ray measurements that have succeeded thus far are those based on the use of photon counters sensitive to comparatively narrow bands of the spectrum, it is still not possible to define the solar spectrum in any but the most approximate terms. In the work of the author and his colleagues at the Naval Research Laboratory, Washington, D.C., we have assumed a Planckian distribution and estimated the total flux on the basis of measurements made in a band from 44 Å to 60 Å. If the coronal temperature was assumed to be 1 million °K, the computed flux was about 0.13 erg cm.$^{-2}$ s^{-1} at solar sunspot minimum in 1953. For an assumed temperature of 2 million °K, the computed flux would have been 0.35 erg cm.$^{-2}$ s^{-1}. To satisfy the requirement that the emission must fall to a negligible flux at 20 Å, as was observed near solar sunspot minimum, the temperature would need to be closer to 0.5×10^6 °K. With the rise in coronal activity approaching solar maximum, more energy was observed to appear at wave lengths below 20 Å.

The shape of the measured distribution in the 8 to 20 Å band, however, required a Planckian temperature of 2 million °K. At that temperature, the Planckian distribution required to supply the observed counting rate of a photon counter sensitive to the 8 to 20 Å band would enhance the intensity of 0.1 erg cm.$^{-2}$ s^{-1} derived from a 0.5×10^6 °K source by less than 1 per cent in the 20 to 100 Å range. It appeared therefore that the sources of the higher temperature emission were restricted to an area of less than 1 per cent of the disk. This is approximately the area of coronal condensations as determined from radio interferometer measurements.

Over a solar cycle, rocket measurements of the total X-ray emission have varied from a minimum of 0.13 erg cm.$^{-2}$ s^{-1} to a maximum of about 1.0 erg cm.$^{-2}$ s^{-1} in the absence of solar flares. In the shortest wave-length emissions, the variations have been much wider. TABLE 1 lists the observed counting rates in the 2 bands, 2 to 8 Å and 8 to 20 Å, from 1949 to 1959. In the 8 to 20 Å band there was a factor of 40 variation from 1953 to 1959, under quiet conditions. In the 2 to 8 Å band, the emission almost disappeared near solar minimum.

TABLE 1

X-RAY COUNTING RATES OBSERVED OVER A SOLAR CYCLE

Time (UT)	Date	Counting rate (counts cm.$^{-2}$ s^{-1})		Solar activity
		2–8 A	8–20 A	
1730	9/29/1949	1.0×10^4		2½ hours after Class 1 flare
1459	6/1/1952	495		Quiet
1344	5/5/1952	<125		Quiet
2240	11/15/1953	<40		Quiet
1546	11/25/1953	332		Quiet
1529	12/1/1953		4.5×10^4	Quiet
2250	10/18/1955		1.4×10^5	Quiet
1915	7/20/1956	1.2×10^5		Late in Class 1
1634	7/24/1959	2.4×10^4	1.0×10^7	Quiet
1600	8/14/1959	1.3×10^4	2.0×10^6	Quiet
2253	8/31/1959	$>7 \times 10^5$	$>1 \times 10^7$	Class 2+

X-Ray Solar Disk Photograph

On April 19, 1960, my associates and I were successful in obtaining the first photograph of the sun in the soft (20 to 60 Å) X-ray region with sufficient resolution to reveal the source distribution of this radiation. The X-ray image is shown in FIGURE 1, together with calcium-K line photographs for the same day, and for the preceding and following days.

The photograph was obtained with a pinhole camera flown in an Aerobee-Hi rocket that reached a peak altitude of 195 km. The pinhole was 0.013 cm. in diameter and the camera was 16 cm. long. Covering the pinhole was a film of Parlodion, a form of cellulose nitrate, which transmitted efficiently from about 20 Å to 60 Å. This served as the substrate for an evaporated

aluminum coating of sufficient thickness (2500 Å) to make the system opaque to visible as well as ultraviolet light. The camera was mounted on a biaxial pointing control, built by the University of Colorado, Boulder, Colo., which kept it directed toward the sun to within less than 1 min. of arc. However, the pointing control did not compensate for motion about the principal axis of the pinhole camera resulting from the rocket's precession. The images of discrete X-ray sources were therefore smeared through about 170° of arc.

Comparison of the X-ray photograph with calcium-K line photographs shows a remarkable correlation when allowance is made for the rotation of

FIGURE 1. (Above) Ca K line photographs of solar disk; (below) X-ray solar disk photograph.

the camera. Two bright plage areas on the west limb, one that is clearly evident a day later on the east limb, and a bright plage in the northeast quadrant account for virtually all of the X-ray emission. Densitometry of the X-ray image indicates that the intensity corresponding to the brightest plage region is about 12 times that from the weakest region of the background. When the effect of the rotation of the camera is taken into account, it appears that the X-ray emission from the brightest plage area may have been at least 70 times as intense as from the quiet background.

Solar X Rays and Decimeter Waves

The slowly varying component of radio decimeter wave emissions appears to originate in coronal condensations at temperatures ranging up to a cutoff

of about 1.6×10^6 °K, with a most typical value being about 0.6×10^6 °K. Waldmeier[1] estimated densities 20 times normal from the observed brightness of the condensations in white light. The radio-emitting regions cover areas equal to the calcium plages and lie above them at heights from several tens of thousands to 100,000 km. These dense regions appear to extend in a columnar form above the photosphere and may be constrained into the shape of streamers by the action of local magnetic fields. Unsold[2] suggests that at a distance of several solar radii nearly all of the material in the corona is confined to approximately one sixth of its volume.

Since both the X-ray emission and the decimeter waves are generated by thermal processes at temperatures of the order of 10^6 °K and, since the intensities of both radiations are proportional to the square of the electron density, it is not surprising that the X-ray disk photograph resembles the plage pattern as closely as does the radio source distribution. The sources of the slowly varying component have lifetimes of the order of 3 months. It has been noted that the E region has a similar 27-day periodicity persisting over several rotations corresponding probably to the lifetime of X-ray emission from a particular coronal condensation.

Solar Flares

The emission of visible light accompanying a moderately large solar flare (Class 2 or 2^+) is of the order of 10^{30} ergs, and most of it is spread over an interval of about one-half hour. Accompanying the visible flash there must be a burst of ionizing radiation capable of enhancing the electron density of the D region of the ionosphere and consequently producing radio fade-out and other sudden ionospheric disturbance (SID) phenomena. Prior to 1956, the favored hypothesis was an enhancement of Lyman-α, but direct measurements from rockets have since revealed no evidence for significant increases of this radiation during the course of a flare. Instead, the observations showed marked enhancement of the X-ray emission, both an over-all increase of the entire X-ray spectrum and an extension to shorter wave lengths. The X-ray emission makes its appearance immediately with the onset of the visible flare and may amount to as much as 10^{31} ergs. It is therefore one of the major features of the flare outburst. In comparison, the energy dissipated in surge ejection of solar gas is of the order of 10^{29} ergs.

A series of experiments in 1957 confirmed the importance of X-ray emission and gave the results listed in TABLE 2. The shortest wave lengths observed were in the neighborhood of 1 to 2 Å and penetrated well below the normal base of D region. The fluxes were high enough to account for the observed SID events. It appeared that the enhancement of X-ray emission could be explained as a consequence of thermal excitation in regions of coronal condensations, with temperatures reaching 10^7 °K. Under such conditions, even atoms as heavy as iron in the coronal gas would be stripped completely of electrons, and the X-ray emission would be expected to reach a short wavelength limit between 1 Å and 2 Å.

The most recent series of measurements conducted by Chubb *et al.*[3] in 1959 demonstrated clearly that X-ray quanta up to energies of at least 125 kev are emitted early in flares of magnitude 2^+ or greater. Quanta in the energy range

30 to 125 kev were detected with a scintillation counter. The instrumentation included pulse-height discrimination and therefore yielded spectral data. TABLE 3 shows comparative results for a quiet sun and for an average Class 2^+ flare about 3 min. after the first visible detection. The flares produced not only very short wave-length X-rays, but the entire spectrum was enhanced. FIGURE 2 shows 2 pulse amplitude spectra obtained during the course of a Class 2^+ flare about 3 min. apart. The spectrum fell rapidly in intensity, about a factor of 2 for every 5 to 6 kev increase in energy, and softened notice-

TABLE 2

X-RAY FLARES

Rocket	Launching time (Z)	Class flare	Altitude of X-ray penetration (km.)	X-ray wavelength (A)	X-ray flux (erg cm.$^{-2}$ s^{-1})
Rockoon NN5.31	7/20/1956 1217	1 No SWF	77	3–8	5×10^{-3}
Nike-Deacon NN7.42F	8/20/1957 0949	1+ weak SWF	70	2.5–?	3×10^{-3}
Nike-Deacon NN7.45F	8/29/1957 1412	2 ½ hr. SWF	77	3–8	2×10^{-2}
Nike-Asp NN7.49F	9/18/1957 1054	Class 3 2 hr. SWF	63.5	1–5–?	2×10^{-5} at 63.5 km. 1.2×10^{-4} at 70 km.

TABLE 3

X-RAY FLUX*

Wave-length interval	Background 8/14/59 1600 UT	2+ Flare 8/31/59 2253 UT
20–100 Å	0.6	4.0
8–20 Å	0.002	0.09
2–8 Å	0.00055	0.03
0.6 Å	—	0.000023

* Erg cm.$^{-2}$ s^{-1}.

ably during the 3-min. interval between the sets of measurements. Although the quantum energies were high, the flux in the short wave-length tail, about 2×10^{-5} erg cm.$^{-2}$ s^{-1}, was a very small portion of the total X-ray flux produced by the flare. The X-ray emission above 30 kev could be described as brehmsstrahlung from electrons thermalized at temperatures in the neighborhood of 10^8 °K.

Lyman-Alpha

Since 1949, many individual measurements of Lyman-α have been made with photon counters, thermoluminescent phosphors, ion chambers, and photocells, and by densitometry of spectrograms. The most reliable results

are believed to be those obtained from ion-chamber measurements. During the past 3 years, the ion-chamber data grouped within the range 3 to 6 ergs cm.$^{-2}$ s^{-1}. One of the early photon counter experiments (1952) measured a flux as low as 0.1 erg cm.$^{-2}$ s^{-1}, and some of the spectrographic results were as low as 0.5 erg cm.$^{-2}$ s^{-1} (1952). If these low values are reliable, they indicate

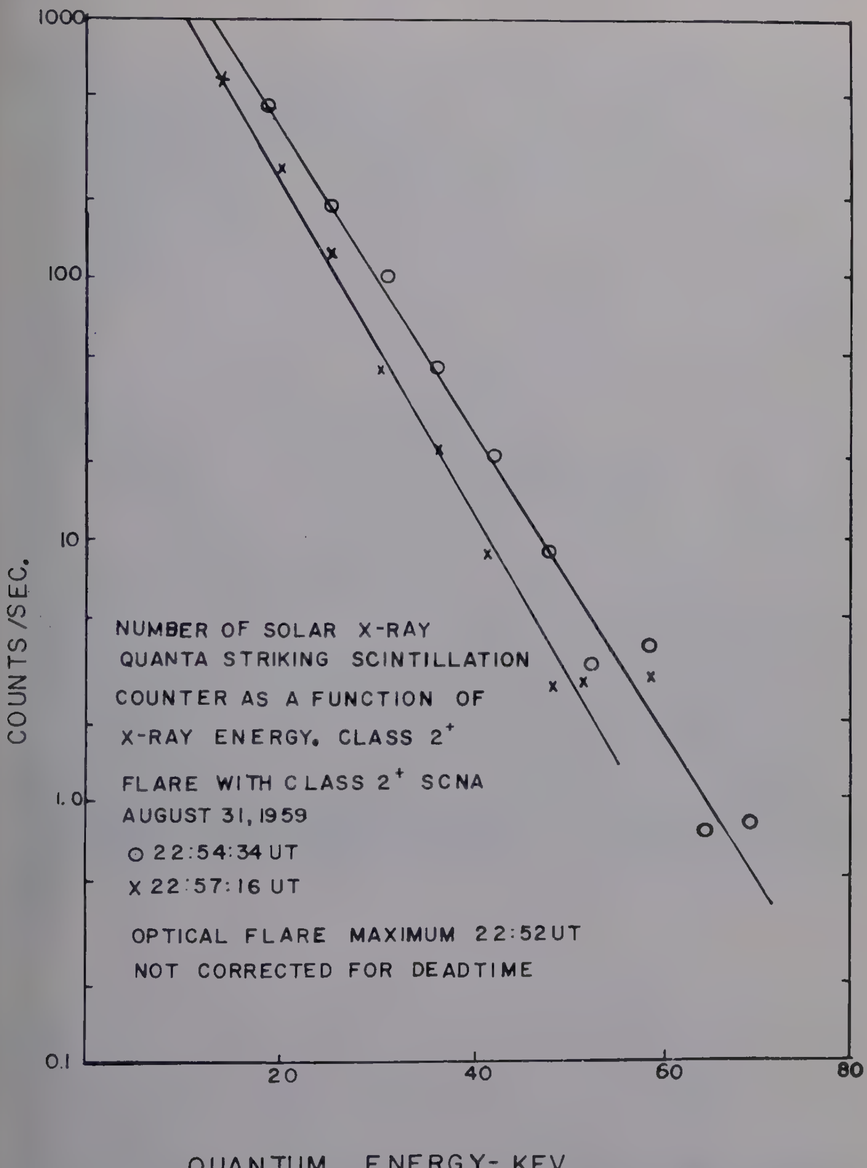


FIGURE 2. Number of solar X-ray quanta striking scintillation counter as function of quantum energy at 2 times during August 31, 1959, flare.

a large solar cycle variation in Lyman-α. Because water vapor is an extremely powerful absorber of Lyman-α, it has been suggested that some of the low-intensity results may have been produced by water vapor carried along with the rocket.

One of the best clues we have to the variability of Lyman-α is a disk photograph (FIGURE 3), obtained by Purcell *et al.*[4] in 1959. It is clear from the

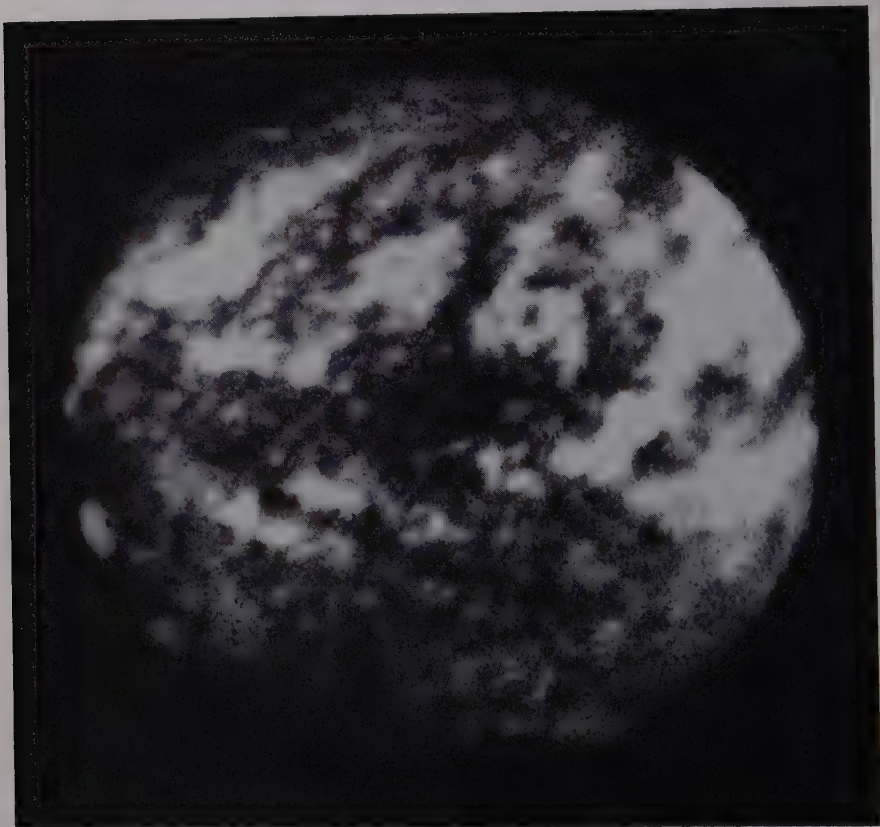


FIGURE 3. Photograph of the sun in the wave length of Lyman-α (1215.7 Å), March 13, 1959.

photograph that most of the Lyman-α emission comes from plage formations. There is a detailed correlation between the picture in Lyman-α made from the rocket and simultaneous photographs made from the ground in Ca-K and H-α. The Lyman-α picture shows a grosser plage structure and much greater contrast than the Ca-K picture. It is well established that the fraction of the solar surface covered by calcium plages reaches a maximum at sunspot maximum and nearly vanishes at sunspot minimum. We may expect Lyman-α plages to follow a similar cycle. In the rocket photograph, the Lyman-α plage regions cover approximately one fourth of the area of the sun, and their intensities

average 5 times the quiet background. Therefore if the plages disappear completely, leaving only the background at solar minimum, the total output of Lyman-α would change by about a factor of 2.

Lyman-α is the principal source of D-region ionization, presumably by virtue of the ionization of nitric oxide that is present as a trace constitutent in the upper atmosphere. X-rays of wave length less than 6 Å can contribute to D-region ionization. Their flux is negligible at solar minimum but appreciable at solar maximum. It is probably true however that the major ionization source for D region, even at solar maximum, is Lyman-α, except for the brief transient effects of solar flares. Radio measurements indicate that the D-region absorption increases by approximately a factor of 3 from solar minimum to maximum. The absorption is proportional to electron density, which varies with the square root of the flux of solar ionizing radiation. Accordingly one might expect the Lyman-α flux to vary as much as a factor of 9, if it is the only variable factor affecting the electron density.

Solar Radiation and Atmospheric Structure

Observations of the variations in drag decelerations of satellites show a strong diurnal variation that can be related only to heating by solar electromagnetic radiations. In the ultraviolet range, from 1500 to 1300 Å, the flux is about 3 erg cm.$^{-2}$ s^{-1}, of which about 0.5 erg cm.$^{-2}$ s^{-1} is immediately converted to heat in the E layer. It is unlikely that radiation in this wave-length range varies significantly over a solar cycle. Also effective in the E region is the absorption of X rays from 10 to 100 Å. The flux varies from about 0.1 to 0.7 erg cm.$^{-2}$ s^{-1} from solar minimum to maximum, and virtually all of this energy goes into heat. At higher altitudes, heating is due to ultraviolet radiation in the wave-length range 200 to 100 Å and, in particular, to the resonance lines of He I and He II. According to Nicolet,[5] a flux of 1 erg cm.$^{-2}$ s^{-1} is sufficient to explain the diurnal variation in atmospheric density revealed by satellite drag measurements.

Priester and Martin[6] and Jacchia[7] have pointed out the close correlation between satellite accelerations and the fluxes of solar decimeter wave radiations. It is well established that the decimeter wave flux varies in almost direct relationship to the areas of calcium plages. The equally close correspondence between X-ray emission, Lyman-α emission, and plage formation implies that variations in atmospheric structure are controlled by X-ray and extreme ultraviolet radiations. We should therefore expect large solar cycle variations in atmospheric density above 100 km. proportional to the variation in solar X-ray and extreme ultraviolet fluxes.

Summary

The total power radiated by the sun is very nearly constant, but the few parts per million emitted in the X-ray and extreme ultraviolet portions of the spectrum are highly variable. Rocket astronomy is now slightly more than 1 decade old, sufficient to indicate the sunspot cycle dependence of the sun's X-ray and far ultraviolet radiations. It seems clear that solar X rays (20 to 100 Å) are a major source of E-region ionization. The X-ray flux rises and falls in the same sense as the coronal green-line intensity, solar decimeter wave

emission, and E-region critical frequency. From 1953 to 1959 the variation was approximately a factor of 600 below 8 Å, a factor of 60 from 8 to 20 Å, and a factor of 7 from 44 to 60 Å. The most marked outbursts of X-ray emissions accompany solar flares and sufficient power is radiated to produce the accompanying ionospheric distrubances. The variations in ultraviolet emission are less clearly indicated by direct measurements because of experimental inaccuracies. However, much can be inferred from solar disk photographs in the Lyman-α line of hydrogen. A variation of the order of a factor of 2 is indicated between solar minimum and maximum.

Acknowledgments

I am indebted to H. D. Babcock, Mt. Wilson Observatory, Mt. Wilson, Calif., and to R. L. Blake and A. Unzicker, United States Naval Research Laboratory, Washington, D.C., for the use of FIGURE 1. I am grateful to R. Tousey, J. D. Purcell, and D. Packer, United States Naval Laboratory, Washington, D.C., for the use of FIGURE 3.

References

1. WALDMEIER, M. 1947. Coronal radiative and ionospheric variations during the solar eclipse July 9, 1945. J. Terr. Mag. Atmos. Elec. **52:** 333.
2. UNSOLD, A. 1960. Aufbau und Variationen der Sonnenkorona. Z. Astrophys. **50:** 57.
3. CHUBB, T. A., H. FRIEDMAN & R. W. KREPLIN. 1960. X-ray emission accompanying solar flares and non-flare sunspot maximum conditions. Proc. 1st Internl. Space Sci. Symp., Nice, France. *In* Space Research. : 695. H. K. Kallmann Bijl, Ed. North-Holland Publ. Co. Amsterdam, Holland.
4. PURCELL, J. D., D. M. PACKER & R. TOUSEY. 1959. Lyman-Alpha photographs of the sun. Nature. **184:** 8.
5. NICOLET, M. 1960. Structure of the thermosphere. Prel. Notes No. 10. Centre Nationale de Recherches de L'Espace.
6. PRIESTER, W. & H. A. MARTIN. 1960. Soläre und tageszeitliche Effekte in der Hochatmosphäre aus Beobachtungen künstlichen Erdsatelliten. Mitteilungen der Universitäts-Sternwarte Bonn. **29**.
7. JACCHIA, L. G. 1959. Solar effects on the acceleration of artificial satellites. Smithsonian Inst. Ast. Obs., Spec. Rept. No. 29.

SOLAR RADIATION IN THE EXTREME ULTRAVIOLET REGION OF THE SPECTRUM AND ITS EFFECT ON THE EARTH'S UPPER ATMOSPHERE*

William A. Rense

Physics Department, University of Colorado, Boulder, Colo.

Introduction

The character of the earth's upper atmosphere is profoundly affected by solar ultraviolet and X-ray radiation. It is the purpose of this paper to review briefly the methods by which such radiation has been studied, the results to date, and the application of these results to the interpretation of the properties of the upper atmosphere. At the present time most of the published data were the result of rocket experiments. A program involving satellite observations is now under way. It need not be mentioned that present knowledge concerning solar short wave radiation is not complete. Nevertheless sufficient data have accumulated to point the way to definite theoretical conclusions.

Instrumentation

I have given a detailed review of instruments used to study ultraviolet radiation in another paper.[1] For present purposes I shall mention three basic kinds of rocket or satellite instruments, namely the photon counter, the grating spectrograph, and the echelle spectrograph. The first of these is especially useful in the detection of X-ray radiation up to about 100 Å. The second is the basic device for investigating the solar spectrum from the X-ray region into the near ultraviolet region. The echelle spectrograph serves the purpose of providing a high-resolution instrument for measuring line profiles and obtaining accurate wave lengths in the ultraviolet region.

A photon counter resembles a Geiger-Müller tube. The important parts of the photon counter are the chamber, the window, the enclosed gas, and the cathode surface. Proper selection of window and gas determines the range of wave lengths to which the tube will respond. Ordinarily, because of the difficulty in finding a window for wave lengths between about 100 Å and 1100 Å, the photon counter is useful in the X-ray region and in the region above 1100 Å. The counter lends itself to quantitative measurements of radiation intensity.

Most grating spectrographs that have been used to study solar ultraviolet radiation have employed the Rowland circle-type mounting. Both normal incidence and grazing incidence arrangements have been employed successfully. These spectrographs resemble those commonly associated with optical laboratories; however, a few modifications have been made in order to overcome difficulties peculiar to rocket and satellite environments. For example, when gratings are used at normal incidence the reflectivity has been increased by the

* The work described in this paper was supported in part by grants from the Geophysics Research Directorate of the Air Force Research Center, Air Research and Development Command, Cambridge, Mass.

use of suitable coatings, such as magnesium fluoride over aluminum. The astigmatism ordinarily present in the grazing-incidence spectrograph has been reduced by the addition of a suitable focusing mirror in front of the slit. Considerable progress has been made in the rulings of gratings, both from the point of view of increasing the intensity of a given order by the proper ruling of a blaze angle and from the point of view of obtaining gratings with short radius of curvature. When photography is the method of detection the most successful film has been the short-wave radiation film of Eastman Kodak Company, Rochester, N.Y. This film is essentially a Schuman type f lm with very little gelatine content. In some cases other forms of detection have been employed, such as the Bendix photomultiplier cell. Such devices are especially useful for obtaining quantitative results and in those cases in which film recovery is not possible and telemetering is available.

An echelle is a reflection grating with a relatively small number of lines ruled per centimeter. It also possesses a strong blaze angle. When used in conjunction with a grating whose dispersion is set at right angles to that of the echelle, the latter becomes a high-resolution element in a spectrograph. An echelle instrument suitably designed for the investigation of ultraviolet radiation can be helpful in obtaining high-resolution profiles of important solar lines, and also in the measuring of accurate wave lengths.

The above instruments, when used in rockets, are not effective unless some means is available for pointing them in the direction of the sun and thus counteract the motion of the rocket. Such a pointing device[2] has been designed and constructed, and has been successfully used in most of the flights that have yielded important data to date.

Results

The ultraviolet region of the solar spectrum has been mapped from about 80 Å up to the 2800-Å region that represents the limit detectable at sea level. More than 1500 Fraunhofer lines and more than 200 emission lines have been recorded and, in many cases, identification of the radiating atom or ion has been made. The general characteristic of the spectrum is as follows. A continuum extends down to about 100 Å or beyond; this continuum is essentially an extension of the one in the visible part of the solar spectrum, although its origin may be different. It is not as intense as the continuum in the visible region, in the sense that the black-body temperature to which it corresponds varies with wave length and is, in general, under 5700° K. In addition to this continuum, the continuum beyond the series limit of the hydrogen Lyman series has been observed, as well as a continuum in the region of the hydrogen Lyman-alpha line itself at 1216 Å. The continuum beyond the ionized helium series corresponding to the Lyman series has also been detected. Fraunhofer lines have been measured down to well below 1600 Å. Hundreds of these have been identified. Solar ultraviolet emission lines make their appearance near 1900 Å. Emission lines appear in fairly large numbers all the way down to below 80 Å.

Because the observations to date have been taken with rockets that do not penetrate the atmosphere much above 215 km., there is still some question as

to the intensity of these emission lines outside the earth's atmosphere. It appears, however, that the Lyman-alpha line due to hydrogen at 1216 Å is the most intense. Its intensity is approximately 6 ergs/cm.²/sec. This value may be assumed to be true for outside the atmosphere because of the low absorption of the atmospheric constitutents for this particular radiation. Another intense line is the corresponding one of ionized helium, namely the 303.8 Å line. At about 200 km. different observers have measured its intensity to be somewhere between $\frac{1}{10}$ and $\frac{8}{10}$ of an erg/cm.²/sec. Opinions differ concerning the intensity of this line outside of the earth's atmosphere. It has been estimated that the total intensity in the emission lines of the solar ultraviolet spectrum is about 10 ergs/cm.²/sec. This figure refers to the regions outside the earth's atmosphere. At the present time it appears that the hydrogen Lyman-alpha line at 1216 Å is the greatest contributor.

Contributions to the intensity of ultraviolet radiation in the continuum mentioned above are not inappreciable. In the X-ray region intensities depend considerably on solar activity, but roughly between 8 and 100 Å the intensity is about 0.1 erg/cm.²/sec.

The origin of the solar ultraviolet line emission is fairly well understood. Most of the lines are due to ionized helium, carbon, oxygen, silicon, and nitrogen, in addition to neutral hydrogen and neutral helium. Most of these lines originate in what is known as the chromosphere of the sun, or at least in the higher parts of the chromosphere where the temperatures are sufficiently high to ionize the atoms. Some of the emission lines are due to highly ionized metallic atoms such as iron, magnesium, and nickel. These originate in the solar corona, where the temperature is of the order of 10^6 K°. The origin of the X-ray radiation is more obscure. Some of it seems to originate as a result of electron accelerations in the corona, and some of it represents optical line emission of highly ionized atoms in the corona and, therefore, is not true X-ray radiation. The origin of the solar ultraviolet continuum that joins the continuum in the visible is not known. Several theories have been proposed. Undoubtedly the explanation lies somewhere within the scope of these theories.

Photographs of the sun's disk in the light of hydrogen Lyman-alpha have been obtained. These showed that the radiation is not emitted uniformly over the entire disk but rather is enhanced in the active regions on the sun's surface, chiefly in the plage areas. Depending upon the presence or absence of the plage areas, hydrogen Lyman-alpha may vary in intensity by a factor of two. Radiation from other atoms, in some cases, also seems to be enhanced in the active regions of the sun. One can expect that the solar ultraviolet radiation will be a variable that depends upon the sunspot cycle. To what extent the variation is significant in so far as the upper atmosphere is concerned is a problem that is being studied at the present time. However, more quantitative data are needed.

The study of the profile on the hydrogen Lyman-alpha line under high resolution has yielded some interesting information concerning the possible presence of hydrogen in the earth's upper atmosphere. The line itself is fairly wide, having a half width of about 1 Å; in the center, however, there is a sharp and deep absorption pit that has been attributed to hydrogen in the earth's atmosphere.

Effects of Solar Ultraviolet Radiation on the Earth's Upper Atmosphere

In a broad sense, the effect of ultraviolet radiation from the sun on the earth's upper atmosphere is to produce dissociation of molecules, ionization, and heating. At the present time no comprehensive explanation of these effects is possible because of lack of accurate data or theoretical parameters. Nevertheless progress has been made in some instances, so that an understanding of the more conspicuous phenomena is nearly complete.

It is now fairly certain that the D-layer in the ionosphere is formed by the action of the Lyman-alpha radiation at 1216 Å on nitric oxide. The NO molecule is ionized by radiation; the NO^+ ions and the electrons so formed represent the ionization in the D-layer. Maximum electron density is at about 80 km. About 20 km. above the maximum of the D-layer, another maximum of ionization appears, locating the so-called E-layer. There is a difference of opinion concerning the cause of the E-layer. However most upper air physicists agree that both X-rays (especially in the range between 15 and 100 Å) and Lyman-beta at 1026 Å are the cause of ionization. It is believed that these radiations ionize molecular oxygen. Some contribution is also made by the ionization of NO. Above the E-layer lies the F-layer. It actually consists of two maxima during the daytime: one locating the so-called F_1-layer at about 150 km. and the other locating the so-called F_2-layer at about 270 km. Here the ionization is due essentially to the effect of the radiation in solar ultraviolet emission lines between 900 Å and 200 Å, and also to the effect of the solar continuum in this region. The strong emission line of ionized helium at 304 Å and the enhanced region of the continuum at the end of the Lyman series are strong contributors. In addition to ionized atomic oxygen, ionized molecular oxygen is an important ion constituent in the F_1-layer. In the F_2-layer ionized oxygen dominates, but perhaps ionized hydrogen is also present along with neutral atoms of these elements.

Unfortunately, conclusive direct evidence concerning the composition of the earth's upper atmosphere is not yet available. However experimenters employing specially designed mass spectrometers are working toward the solution of this problem. Some idea of the composition may be obtained as a result of the interplay of theory and incomplete observations. Below the D-layer, ozone makes its appearance as a result of the action on molecular oxygen of the solar ultraviolet radiation between 1925 Å and 1760 Å; ozone absorbs heavily in the 2900 to 2200 Å region. Molecular oxygen begins to dissociate at an altitude of about 100 km. (dissociation, however, is not complete until a much higher altitude is reached; the latter may be in the F-layer). Dissociation is accomplished by ultraviolet radiation of wave lengths less than 1760 Å.

The ionization processes that occur in the F-layer involve radiation that carries energy in excess of that required for ionization. The extra energy is given, in part, to the freed electrons as kinetic energy. The electron temperature is thus raised, with the result that the kinetic temperature of the F-layer tends to rise. The heating effect would depend upon the intensity of the incident flux and the efficiency of the heating process. The temperatures achieved and the temperature gradient established will be determined largely by the method of transference of heat to the surroundings. Careful considera-

tion of all the processes involved point to the likelihood that conduction is the chief means of carrying off the internal energy given to the gases of the F-layer by the ionization processes. Calculations made on the assumption of an upper atmosphere composed of atomic oxygen lead to observed values of temperatures and temperature gradients when the known data on flux and acceptable values of thermal efficiency are substituted in the resulting equations.[3] At very high levels, hydrogen and, perhaps, helium make up most of the atmosphere; these will be largely in the ionized state, especially at higher levels. The presence of neutral hydrogen atoms in the earth's atmosphere seems to have been substantiated by recent rocket data.

What is the effect of solar activity on the solar ultraviolet spectrum, and what effect do variations in the latter have on the earth's upper atmosphere? First, it has been shown conclusively that considerable variations in X-ray emission from the sun take place following, in general, the solar cycle and, in particular, certain types of disturbances on the disk of the sun, especially flares. The flux may change by one order of magnitude, with the consequence that ionization, especially in the E-layer, may appreciably change. Evidence concerning the fluctuation of the solar ultraviolet intensity tied up in the emission lines between 1900 Å and 15 Å is not conclusive. Nevertheless it has been established that the intensity of the radiation in some cases is sensitive to the region of origin on the sun's disk. Hydrogen Lyman-alpha, for example, is enhanced in the plage areas in the neighborhood of sunspots. Other emission lines are also associated with plage areas and still others with bright coronal regions. Because these special regions of origins of the lines indicate solar disturbances, one can predict with certainty some variation in the intensity of solar ultraviolet emission line spectra. The variation however, is not as great as that for X rays. Nevertheless, because considerably more energy is tied up in the former radiation, the effects may be felt in the upper atmosphere. To date there are no data that establish this fact for rays other than X rays. The solar ultraviolet continuum may also vary with the solar activity. Observations have not been complete enough to establish this possibility. There appears to be little variation in the ultraviolet continuum between 2800 Å and 1500 Å. This radiation is especially important in the establishment of the ozone layer and in the dissociation of molecular oxygen. One would expect, therefore, very little connection with the solar cycle as far as these aspects of the upper atmosphere are concerned. Little is known about the continuum below about 1500 Å, and more observations are needed to establish its intensity, not to mention variation in solar activity. The origin of the continuum is also obscure. The near ultraviolet continuum, which seems to show little variation with the sunspot cycle, probably originates in the photosphere or lower chromosphere. The far ultraviolet continuum may originate as a result of the blending of continua formed at the end of series limits of the spectra of various ionized atoms in the chromosphere and corona. Indeed an enhancement at the series limit for the Lyman series of both hydrogen and ionized helium is clearly observed. Should this reasoning be true, at least in part, then one may expect a variation in continuum intensity with solar activity in a manner similar to that of the emission lines in the far ultraviolet. Since the intensity in the

continuum is about equal to that in the emission lines the effect on the earth's upper atmosphere may be significant.

References

1. Rense, W. A. 1960. Recent advances in space solar observatory instrumentation A.P.S. Journal. : 313–322.
2. Wilshusen, F. 1960. Biaxial Pointing Control. Final Report, Contr. No. AF 19(604)-1724 (TN-60-685).
3. Hunt, D. & T. Van Zandt. 1961. Critical points in F layer density and temperature distribution. Research to be published.

THEORY OF JUPITER'S DECAMETRIC RADIO EMISSION

James W. Warwick
High Altitude Observatory of the University of Colorado, Boulder, Colo.

In this paper I shall present what I believe are the first dynamic, that is continuous, recordings of the spectrum of Jupiter's sporadic radio emission, in the long wave-length range, from 15 to 34 mc./sec. Since the discovery of these emissions by Burke and Franklin (1955), their detailed description has advanced rather slowly, after an initial quite active period of research in the interval 1955 to 1957. The number of observatories attempting to observe the sporadic emissions is small, even though our knowledge would be enhanced greatly by synoptic records kept over continuous periods. The emissions, intense by the standards of radio astronomy, occur in a spectral range in which communications signals are enormously strong and the shielding or absorption effects of the ionosphere begin to be important for external signals. Both of these effects have tended to become more severe as the sunspot cycle progressed toward its maximum in early 1958. Finally, Jupiter itself, in an unfortunate coincidence revolves around the sun in phase with the sunspot cycle, and has been in the southern part of the zodiac, close to the ionospheric cutoff for observatories in the northern hemisphere during the sunspot maximum.

Wild and Sheridan's (1958) direction-finding technique, which incorporates phase switching in a swept-frequency interferometer, provides a useful increase in the sensitivity for our dynamic spectrograph. The ultimate sensitivity of the equipment follows from the intensity fluctuations of total power in the antenna beams of the two antennas. With a gain of about 9 db. over an isotropic radiator, the radio source *Cassiopeia A* stands out clearly above the fluctuations of total power. Since our sweep rate is nominally 0.29 mc./sec., repeated at 1.30 sec. intervals, Jupiter emissions should be recorded if they last longer than about 2 sec. at an intensity greater than about 5×10^{-22} flux density units (Wells, 1958). In order to achieve this high sensitivity, we integrate in the frequency domain, by an amount corresponding to about 1 mc./sec. Fine structure in frequency, on a scale smaller than 1 mc./sec. will therefore be smeared out. Nevertheless, features with broader bandwidths often appear.

The antennas are fully steerable corner reflectors, permitting us to track Jupiter during its nighttime appearance of 1960, from 28 January to 1 October. Severe ionospheric interference as well as the low elevation of the source do not permit observations when Jupiter is near (superior) conjunction. Even our early evening records, from August and September, are incomplete and difficult to assess.

Robert H. Lee, of the High Altitude Observatory, designed and built the receiver. It is a double-conversion superheterodyne, employing 2 tracked RF circuits and 2 IF circuits at 7 mc./sec., followed by a conventional second IF strip employing a mechanical filter of 16 kc./sec. bandwidth. The tracking of

the RF stages is optimized by cam-driven, slug-tuned inductances, used as trimmers in addition to conventional rotating, split-stator condensers. The second detector is followed by highly unconventional "minimum detection" circuits (Lee, 1957) designed to reduce blanketing effects of station interference on the final, intensity-modulated display.

The Magnitude and Direction of Jupiter's Dipole Moment

Drake and Hvatum (1959) recently suggested that Jupiter's decimeter radiation originates from synchrotron radiation by electrons trapped in Jupiter's magnetic field. The suggestion was strongly supported, of course, by analogue to the earth's van Allen belts. Field (1959, 1960) systematically reviewed the observations available to him and concluded that nonrelativistic electrons in a strong magnetic field must produce the decimetric radiation by gyroradiation. Field's principal argument against the synchrotron mechanism was that it requires many times more electrons of a given energy in the Jupiter belts than there are in the earth's. Radhakrishnan and Roberts (1960) directly observed the source size at 960 mc./sec., and found the emission coming from regions about $3R_{♃}$ in extent in the E-W direction. This observation does not constitute a firm disproof of Field's interpretation, although with the new observations, the polar field strength of Jupiter must be even larger than the large value he assumed, now becoming about 10^4 gauss.

Observations of the polarization of the decimeter source as a function of frequency offer, as Field has shown, a satisfactory way to settle the issue between Drake and Field. The measures by Radhakrishnan and Roberts indicated 30 per cent plane polarization at 960 mc./sec., with the E-vector parallel to the planet's equator. Unfortunately, possible ionospheric rotation of the plane in the earth's magnetic field makes interpretation not quite as straightforward as would be desired.* One would wish to determine, for example, any obliquity of the dipole moment to Jupiter's rotational axis by observing the rocking of the polarization plane as Jupiter rotates. Also, with high resolution interferometry, it might be possible to set upper limits to the decentering of the moment in two coordinates, above the planet's equatorial plane and from the axis of rotation.

If the synchrotron mechanism is the principal source of the nonthermal decimeter radiation, we can derive a numerical value for Jupiter's dipole moment from Radhakrishnan and Roberts' measurement. We suppose that the diameter of the decimeter source equals the diameter of Jupiter's radiation belts, a likely circumstance in the case of synchrotron emission, but not necessarily valid for gyro emission. Analogous to the diameter of the earth's belts, the diameter of Jupiter's belts is set by the interaction zone between the magnetic field of the planet and the interplanetary gases of solar origin. Since Jupiter is immersed in interplanetary gases presumably like those affecting the earth, the magnetic field, H_0, in the interaction region must be about the same as in the earth's outer van Allen belt. This value, $H_0 \sim 10^{-3}$ gauss, allows us to determine Jupiter's dipole moment, $M_{♃}$, in terms of the radius, R_0, of Jupiter's decimeter source.

$$M_{♃} \sim H_0 R_0^3; \tag{1}$$

* My thanks are due to George Field and James Roberts for pointing this out to me.

$R_{♃} = 7.135 \times 10^9$ cm., Jupiter's equatorial radius. At 960 mc./sec., the measured diameter of the decimeter source is about 3 $R_{♃}$. If we take $R_0 \sim 6\ R_{♃}$, to allow for extension beyond the limits of radio observability, then $M_{♃} = 8 \times 10^{28}$ gauss cm.^{3}

Although the decimeter observations do not provide completely satisfactory evidence, we shall assume that $M_{♃}$ lies parallel to the rotational axis.

The Location of the Moment within Jupiter

The symmetry of the low-frequency source suggests some important conclusions as to the position of the moment within the planet. Many of our conclusions can be checked directly when more complete observations of the decimeter source are available. For example the position of the emission center of gravity should define the position of the dipole; the direction of the polarization, the direction of the moment.

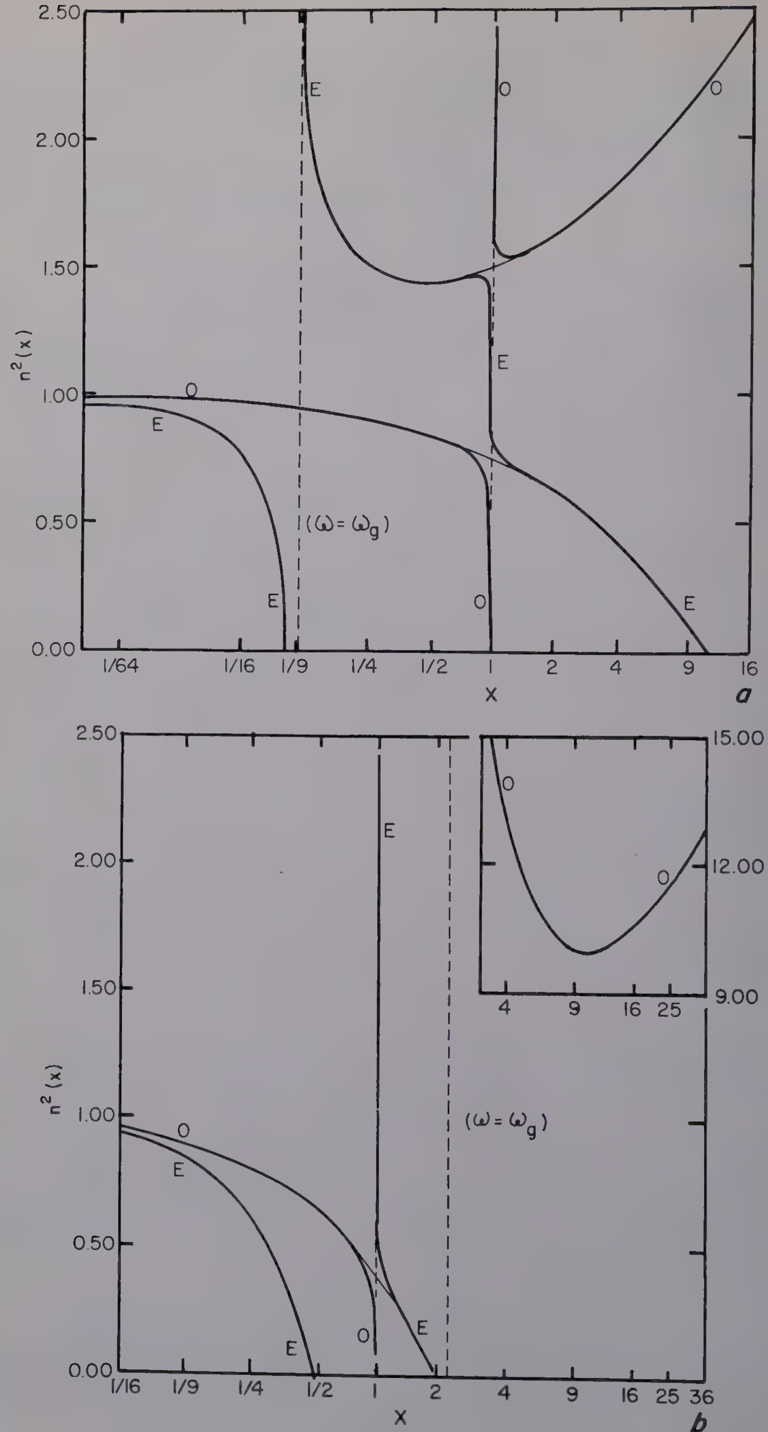
According to Gallet (reported by Carr, 1959) the decameter emission occurs when a shock wave generated at the solid surface of Jupiter passes out through the ionosphere and excites plasma oscillations on its way. Explosions, possibly chemical (Sagan and Miller, 1960) or volcanic (Gallet), recur at activity centers distributed nonuniformly in longitude on the surface of the planet. Since the decameter polarization is virtually always right-handed, we conclude that the volcanoes must be restricted to one or the other of Jupiter's magnetic hemispheres, where the magnetic field is of one sign. Since the longitude profile shows the sources within about one meridional hemisphere, and the polarization shows them to be within one polar hemisphere, they cannot cover more than one quarter of Jupiter's surface. The basis of the internal hypothesis is a highly lopsided distribution of unusual activity centers.

The hypothesis of our interpretation is that the decametric emission is excited by interaction of externally incident particles with Jupiter's ionosphere (Warwick, 1960). The principal reason for supposing an external mechanism is the correlation of emission with solar activity (Warwick, 1960; Carr *et al.*, 1960). It is to be noted that a positive correlation must also be accounted for in the internal hypothesis. In the external hypothesis, we consider that the asymmetry in longitude and polarization result from the decentered location of the planet's dipole. A centered oblique dipole, tipped with respect to the rotational axis seems to be excluded because it will produce equal amounts of right-handed and left-handed polarization, and will show two maxima when it interacts with external particles during each rotation of the planet. The decimeter source indicates that the dipole lies parallel to the rotation axis.

It is to be concluded that the planet's dipole is to one side of the rotation axis, and above, or below, the equatorial plane. This assumption plays the same role in the external hypothesis as the lopsided distribution of volcanoes does in the internal hypothesis.

A Mechanism for Excitation of Jupiter's Decameter Emission

There seems to be some possibility that Jupiter's ionosphere may not have enough electrons to sustain plasma oscillations at 20 mc./sec. (Field, 1959), although others conclude that it can (Rishbeth, 1960; Zhelezniakov, 1958). We note that Zhelezniakov interpreted the experimental results completely in



2.50
2.00
1.50
1.00
0.50
0.00
$n^2(x)$
E
O
O
E
O
E
E
E
O
($\omega = \omega_g$)
1/64
1/16
1/9
1/4
1/2
1
2
4
9
16
X
a
2.50
2.00
1.50
1.00
0.50
0.00
15.00
12.00
9.00
$n^2(x)$
O
O
E
O
E
E
O
E
($\omega = \omega_g$)
4
9
16
25
1/16
1/9
1/4
1/2
1
2
4
9
16
25
36
X
b

FIGURE 1.

terms of the millisecond burst structures (Kraus, 1958) although not all observers confirm their existence (Gardner and Shain, 1958). In any event, the most frequent emissions reported by all observers, the 1-to-10 second bursts or burst complexes, cannot easily fit into the plasma relaxation process suggested by Zhelezniakov. Franklin and Burke (1958) and Gardner and Shain (1958) observed that emission periods were longer at low frequencies (typically 18 mc./sec.) compared to high frequencies (26 mc./sec.). Close to the plasma frequency, the emission should behave in exactly the opposite way, leading at least one worker in the field to question seriously the hypothesis that Jupiter's ionosphere does cut off the decameter waves (Peek, 1959). Initially, Gallet inferred the existence of an ionospheric cut-off on Jupiter from the decline of sporadic emission with increasing solar activity on the upswing into the recent sunspot maximum. I suggest that this phenomenon refers to the changing properties of the interplanetary medium as a function of time in the sunspot cycle (Warwick, 1961).

It appears quite possible that Jupiter's ionosphere is effectively underdense for the decameter emission. Under what circumstances can emission be excited by particles impinging on a low-density plasma or conversely, if the plasma is overdense, can the radiation essentially originate above the critical level? We suggest that the basic phenomenon is Čerenkov radiation, suggested by L. Marshall (1956) and developed extensively by Ginzburg and Zhelezniakov (1959) for the case of solar radio emission. Individual particles excite this radiation when their velocity exceeds the phase velocity of the radiation in Jupiter's ionosphere. The radiation lies in one of the characteristic modes of the magnetoionic medium, and travels essentially in the direction of the exciting particles. The wave that is most strongly generated has a phase velocity near the particle velocity (Denisse, 1960). Pines and Bohm (1952) showed that the intensity of plasma waves excited in this way varies directly as the electron density of the medium.

Introduce the magnetoionic variables $x = \omega_p^2/\omega^2$ and $y = \omega_g/\omega$. Here ω_p is the plasma frequency, ω the wave frequency, and ω_g the gyro frequency, all in circular units. The index of refraction n^2 of the various characteristic wave modes can be conveniently plotted on a graph of n^2 in terms of x. The parameter y for a wave of a given frequency essentially does not vary through the ionosphere, but the variation of n^2 with x depends critically on the assumed value of y. Since only waves for which $n^2 > 1$ can be generated, we plot the dispersion curves, $n^2(\omega^2)$, for the medium. We wish to compare the resultant curves with the plot of $n^2(x)$, and so have actually graphed $n^2(1/\omega^2)$ in FIGURE 1. These curves assume that there are no collisional effects. They show the dispersion for the two cases $\omega_p \gg \omega_g$ and $\omega_p \ll \omega_g$. A horizontal region near some value $n^2 > 1$ defines the velocities of the incoming particles and of the waves that they excite. The corresponding frequency values, where the dispersion curves intersect the horizontal region, represent the frequencies of the

FIGURE 1. (*a*) Magnetoionic dispersion curves for the case $\omega_g > \omega_p$, actually plotted for $\omega_g = 3\omega_p$. At the plasma frequency, $x = 1$, the curves branch, as also at the gyro frequency, $x = 1/9$. The mode is indicated on each branch, E for extraordinary, and O for ordinary. These curves assume nearly longitudinal propagation. (*b*) As in *a*, but for the case $\omega_g < \omega_p$. The plot is for $\omega_g = 2/3\,\omega_p$. Note the change in scale of n^2 for the ordinary branch at $x > 9/4$.

waves that can be excited by the incident particles. FIGURE 1 is sketched for the case of almost-longitudinal propagation; the singularity of these curves at $\omega \sim \omega_p$ disappears in the strictly longitudinal case.

Note that three frequencies are excited for any $n^2 > 1$. When $\omega_p \ll \omega_g$, the particles excite the extraordinary mode (Bremmer, 1949) slightly below the gyro frequency, slightly above the plasma frequency, and at an intermediary frequency, for n^2 less than a certain value. If $n^2 >$ this value, the extraordinary mode is excited below the gyro frequency, and the ordinary mode is excited at two frequencies below the plasma frequency. For the case $\omega_p \gg \omega_g$, two values below the gyro frequency are excited in the ordinary mode, and one value slightly below the plasma frequency is excited in the extraordinary mode.

Not all of these waves escape. To illustrate this point, FIGURE 2 exhibits, under conditions similar to those shown in FIGURE 1, the behavior of $n^2(x)$, for two conditions, $y \gtrless 1$, corresponding to abscissas to the right and left, respectively, of the singularity at $1/\omega_g^2$ in FIGURE 1. Under the conditions of ray optics, waves that are generated on one of the curves of FIGURE 2 will always propagate along the same or similar curves.* The variation of x through the atmosphere depends on the variation of ω_p. FIGURE $2a$ and b show that waves do not escape that are generated in the ordinary mode below the gyro frequency or in the extraordinary mode above the gyro frequency. The ordinary mode does not exist above the gyro frequency for $n^2 > 1$. Therefore Čerenkov radiation may be generated and escape from the ionosphere only in the extraordinary mode below the gyro frequency. This radiation will appear either slightly below the gyro frequency, above the plasma frequency, or at an intermediary frequency, if the velocity of impinging particles is high, but if the incoming particles are slow the radiation will appear only at one frequency, slightly below the gyro frequency. Also, $\omega_g > \omega_p$ at the point of generation of the radiation.

The electron density peaks at a critical value determined either by photoionization from sunlight, or by collisional ionization produced by the fast particles. The majority of radiation originates, all things being equal, at the point of greatest plasma density. If the ionosphere is underdense, only mirrored electrons moving outward through Jupiter's ionosphere can produce the observed radiation. In this case, incident electrons excite waves moving downward, and these pass on through the ionosphere to the solid surface of the planet. For an overdense ionosphere, the radiation lies in the mode reflected at the $1 + y$ turning point, and thus reappears although initially generated downwards.

In FIGURE 3 is indicated, on an $n^2(x)$ plot, a horizontal region representing the velocity of impinging particles, and a vertical region at the maximum x, the critical value x_c. The frequency of the wave follows from FIGURE 1. An escaping mode crosses the intersection region only if this $x \lesssim 1$, as drawn in the figure. Since $n^2 \gg 1$, $y = 1 - \epsilon$, where $\epsilon \ll 1$, for the mode passing through the

* In fact, these curves depend on the angle of propagation of a wave to the magnetic field. This value changes along a given ray path; however, the topology of the curves remains the same except at the singular point of longitudinal propagation. When collisions are unimportant essentially no energy propagates longitudinally.

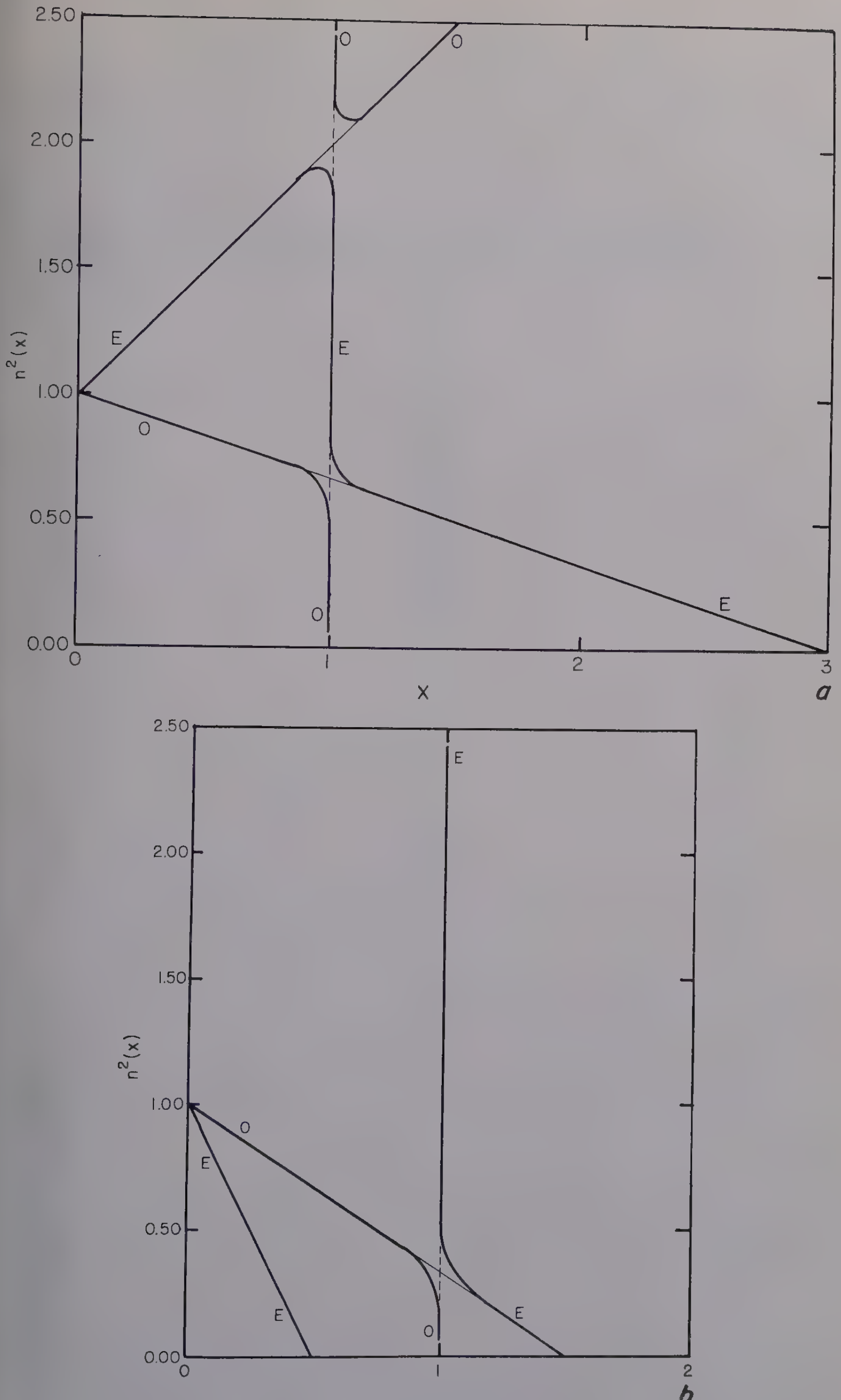


FIGURE 2. (*a* and *b*) The relation between n^2 and x for a fixed value of $y = \omega_0/\omega$. In *a*, $y > 1$, in fact, $y = 2$, while in *b*, $y < 1 = \frac{1}{2}$. As in FIGURE 1, the curves are drawn for nearly longitudinal propagation.

intersection. If $x_c > 1$, the most intense waves that escape are generated slightly to the left of $x = 1$.

Pines and Bohm (1952) give the intensity of the *plasma* wave excited by fast

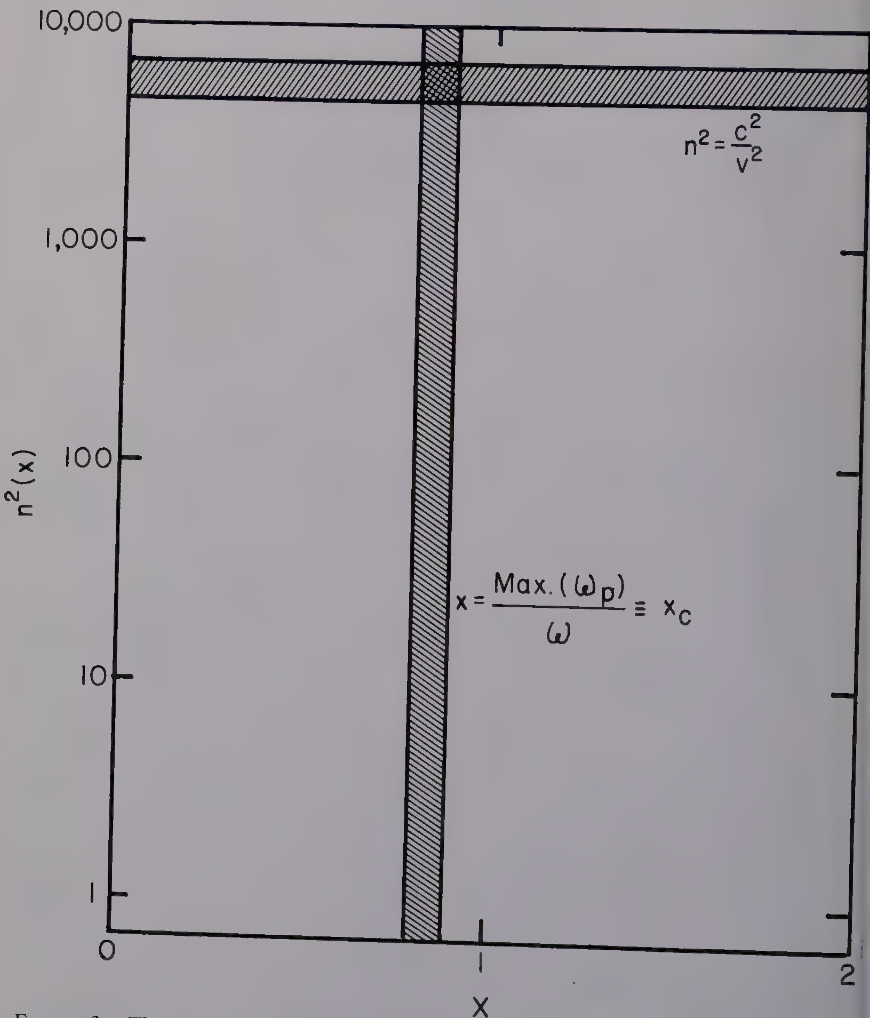


FIGURE 3. The condition for generation of a Čerenkov electromagnetic wave is that incident particles at velocity v correspond to an allowed index of refraction n. The most intense waves arise at the level of maximum electron density or, if this value is too high, in any case just below the level $x = 1$. The intersection region defines a point on a curve of the type shown in FIGURE 2a, the E mode $x < 1$.

particles in the Čerenkov process. We apply this formula also to the case of excitation of electromagnetic modes, where the difference in the index of refraction between plasma and electromagnetic waves is not great. Note that the dispersion curves in FIGURE 2 represent the approximate curves also for the case of a plasma wave. In this case, the singularity at $x \sim 1$ becomes

essentially part of the dispersion relation for a plasma wave, for example, the relation $n^2 = (c^2/v_0^2)(1 - x)$, where v_0 is the thermal velocity of the electrons in the ionosphere. If $c^2/v_0^2 \gg 1$, this relation essentially is a vertical line near $x \sim 1$. In other words, the introduction of thermal motions of the ionospheric electrons leads to a third characteristic mode which, however, does not radically change the nature of the electromagnetic modes near $x \sim 1$. The resemblance of the curves suggests the appropriateness of using the Pines and Bohm relation to compute the excitation of plasma waves. This relation is:

$$\frac{dE}{dz} \sim \frac{1}{4} \frac{\omega_p^2}{E_0} me^2 \log \left\{1 + \frac{E_0}{kT}\right\} \tag{2}$$

in which dE/dz is the energy per unit path length given up by a particle of energy E_0 to the excitation of plasma oscillations; m and e are the mass and charge of the electron, respectively, while T is the temperature of the ionospheric electrons. Equation **2** holds strictly for individual particles only, but we shall apply it to a particle flux on the basis that cooperative phenomena will in all likelihood increase the intensity of the emission rather than diminish it. We shall find that the decameter emission requires a much greater density of electrons than are present in the terrestrial auroras.

Letting $\omega_p = 2\pi$ (10 mc./sec.), and taking $T = 140°$ K, and $E_0 = 10$ kev, we find $dE/dz = 1.71 \times 10^{-22}$ erg·cm.$^{-1}$. The total radio power in low-frequency emission, into all directions, is typically $\sim 5 \times 10^{16}$ erg·sec.$^{-1}$. If the flux of particles, cm.$^{-2}$·sec.$^{-1}$, is F onto Jupiter's surface element dS, and the path length over which emission takes place is L, then

$$L \frac{dE}{dz} \int_{♃} F\, dS = 5 \times 10^{16} \text{ erg}\cdot\text{sec.}^{-1} \tag{3}$$

The surface area of Jupiter is 6.40×10^{20} cm.^{2}, so that the average flux over the *entire* surface of the planet is 5×10^{10} electrons cm.$^{-2}$·sec.$^{-1}$, if we assume that the electrons excite electromagnetic waves over a distance $L =$ 100 km. Fluxes of 10^{11} electrons cm.$^{-2}$·sec.$^{-1}$ often occur in the earth's auroral zones. However, the earth's zones cover perhaps only 10^{-4} of its surface. If we divide the average flux over the surface of Jupiter by this figure, we estimate a flux of 5×10^{14} electrons cm.$^{-2}$·sec.$^{-1}$, required to produce the decameter radio emission in Jupiter's auroral zones. The actual density of the fast particles is 8×10^4 cm.$^{-3}$, much less than the density of the photoelectrically-produced electrons in Jupiter's ionosphere. Such a large flux may produce sufficient ionization so as to justify the assumption of an overdense ionosphere on Jupiter. In that case we feel that the waves are generated at about the level where $\omega_g \gtrsim \omega_p$, and are reflected at $x = 1 + y$.

Assume that electron energies above 1 Mev are required to produce Jupiter's decimeter source by synchrotron emission. We use the convenient tabulation by Dyce and Nakada (1959) of the density of electrons in the earth's inner belt above this energy, to find 10^{-4} electrons cm.$^{-3}$. Scaling up by 10^3, we find 10^{-1} relativistic electrons cm.$^{-3}$ in Jupiter's belt. It would appear from the formula given by Field (1959) that these electrons can account for the decimeter source.

We conclude that to excite the decameter or decimeter emissions requires a much greater density of electrons than are present in the earth's belts, or

auroras, by a factor greater than 10^3. The *same* factor will account for both emissions in a surface magnetic field of the order of 10 gauss.

Drifts of the Decametric Emission Frequency in Time

During the period January 28, 1960, to September 30, 1960, the Boulder radio spectrograph tracked Jupiter continuously during the night. Through July 31, 1960, we recorded emission on 36 separate nights. Until May 1, 1960, the spectra were recorded at a sweep rate of 13.6 (mc./sec.) $sec.^{-1}$ and, afterward, at 27.2 (mc./sec.) $sec.^{-1}$. The limiting sensitivity equals roughly the flux density of Cas A at 18 mc./sec., but the minimum detection scheme employed to reject interfering stations reduces the chances that we can observe the millisecond burst structures reported by Kraus (1958).

Nevertheless, the slower burst structures often appeared very strongly in our recordings. FIGURES 4 and 5 present examples of strong emission as observed with the phase-switching interferometer. The original records were made on facsimile paper, with the frequency range covering 9 inches. The resolution of the paper permits the separation of strong signals about 0.5 mm. apart in the frequency direction, and of successive traces in the time direction about 0.1 mm. apart, as long as the intensity is not too great. These reproductions do not show the resolution limit in the time direction. In general, individual bursts less than several seconds long do not show as distinct phenomena on our records. The successive traces tend to become integrated with one another when they are viewed as in FIGURES 4 and 5. This smearing effect makes it difficult to identify definite features on one trace alone, unless the feature is quite strong. Also, there do appear structures lasting 5 or more sec. that may be the groups or burst complexes described by Gardner and Shain (1958). These structures often reach high enough levels to obscure shorter variations of possibly smaller amplitude. In agreement with frequently-described results of lack of correlation between adjacent fixed-frequency records, on May 11, 1960, fixed-frequency records within 2 mc./sec. of one another would have shown no detailed correlation.

Several striking features of our records do not seem to have been previously described. On many occasions we have observed a general, broadband structure, covering 5 or more mc./sec. and persisting for intervals greater than 1 hour. There frequently occur slow drifts of emission from one range to another. On February 4, 1960, for example (see FIGURES 4 and 5) emission appeared first at the lower limit of our frequency range, 15 mc./sec. and evolved, over a period of about 1 hour, to the upper end, 34 mc./sec. During this emission there was occasional burst activity within the over-all structure, as well as considerable gross structure in the development pattern. At 34 mc./sec. incidentally, this is the highest frequency at which the sporadic Jupiter emission has been recorded, and tends to confirm Kraus' (1958) report of an audio observation at 43 mc./sec. The event of February 4, 1960 lasts a shorter time at the high end of the frequency range than at the lower end, and may be like Franklin and Burke's (1958) and Gardner and Shain's (1958) observations of shorter duration at higher frequency.

The drift in frequency has occurred rather often during the interval of our observations. Remarkably, however, the *sense* of the drift is not always that

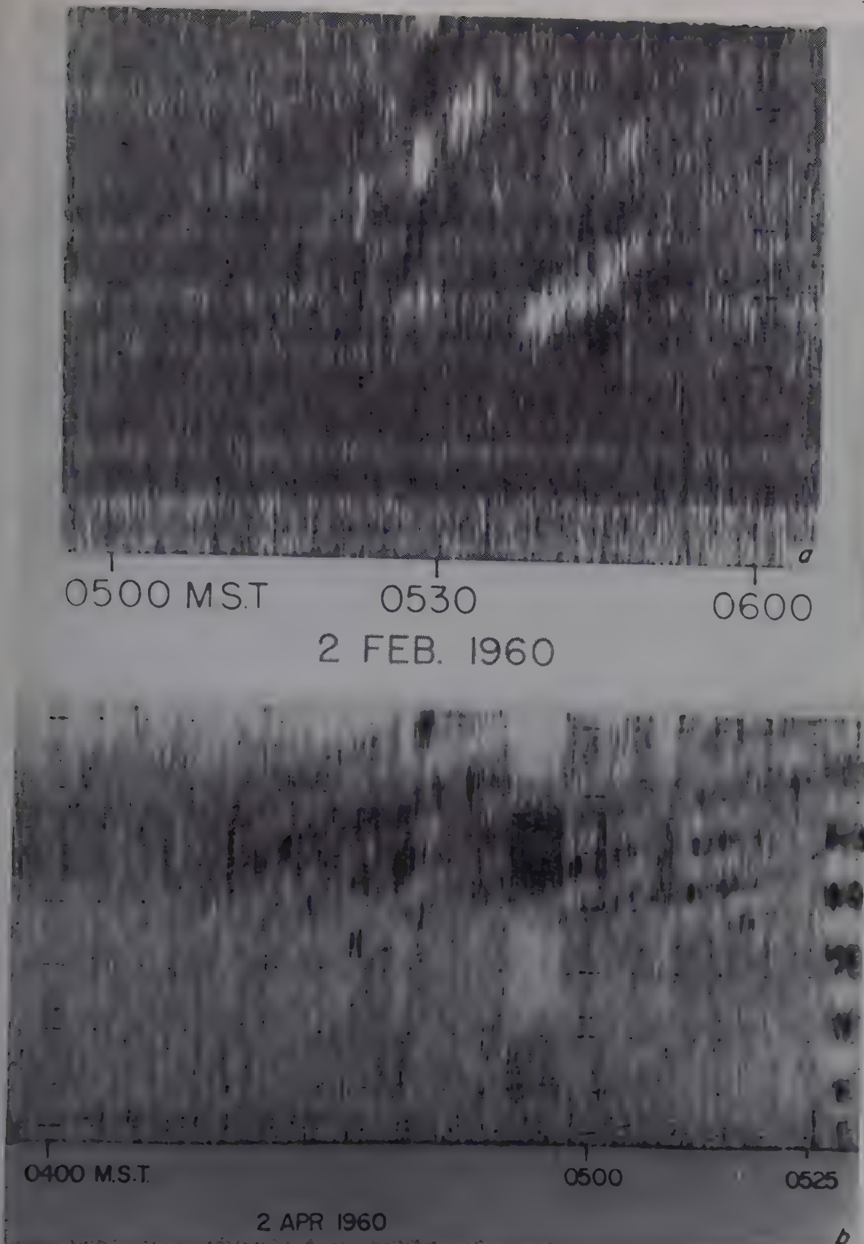


FIGURE 4. (*a* and *b*) Examples of dynamic spectra of Jupiter's decametric emission. Frequency varies linearly from 15 mc/s (*top*) to 34 mc/s (*bottom*). The interferometer fringes appear as diagonal streaks cutting across the spectra. Both *a* and *b* occurred near 200° long., and represent characteristic spectra for that longitude. Note in *b* that a strong source of interference is superimposed on Jupiter's emission from 0451 to 0457 MST (105° W).

of the February 4, 1960 event, but can be opposite, for example, like the event of May 11, 1960 (FIGURES 4 and 5). The latter event also shows a peculiar tendency towards *doubling*, an emission period paralleling the principal emission in its drift, but occurring at a different time. If the dynamic spectrum of the

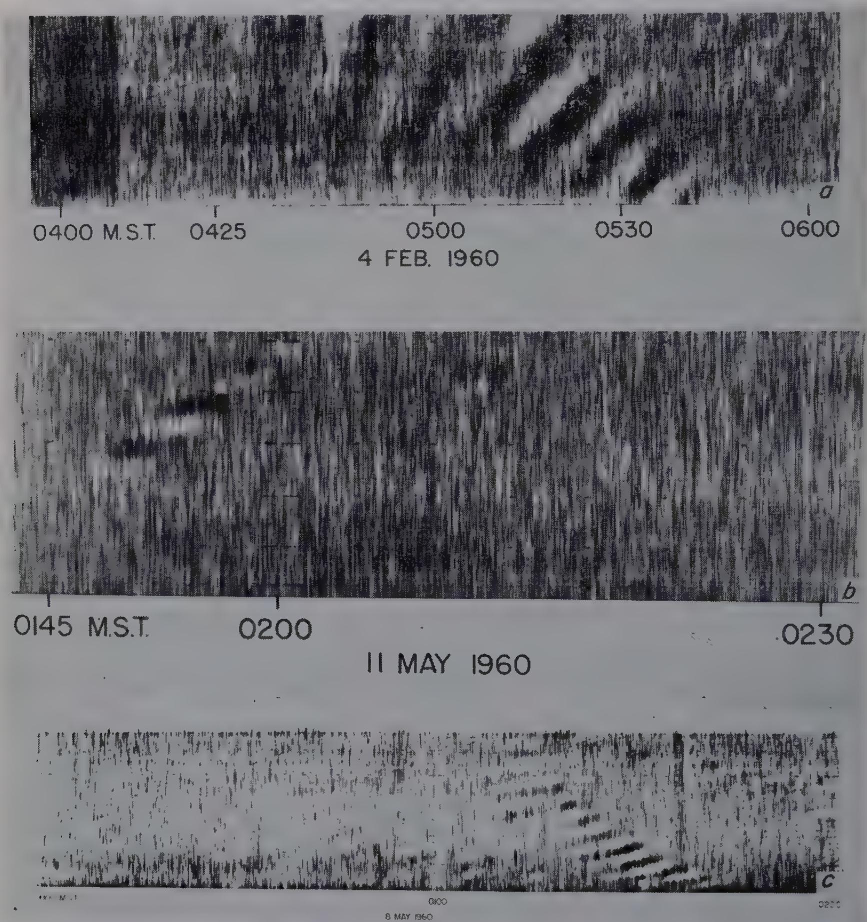


FIGURE 5. (*a* and *b*) More examples of spectra, from positions early (*a*) and late (*b*) in the longitude profile. Note the pronounced drift in frequency of the peak emission. The emission from early longitudes increases in frequency with time, and emission from late longitudes decreases. (*c*) Another spectrum, early in the longitude profile, showing the characteristic positive drift of emission from this position on the planet.

later emission is extrapolated to the earlier time, the frequencies in the two bands are roughly in harmonic ratio. TABLE 1 lists the occasions on which we have found drifts, the longitudes of the drifts,* the sense of the drift, positive

* On a modified System III (Carr *et al.*, 1958) in which the numerical values agree with System II on 1.0 January, 1957, as in System III, but with the period $9^h\ 55^m\ 29^s.5$ (Gallet, 1958) instead of the $9^h\ 55^m\ 28^s.8$ value of System III.

if frequency increases in time and negative if it decreases, the drift rate, and a qualitative statement as to whether the emission shows doubled structure as did the May 11, 1960 event.

We note, in TABLE 1, that positive-drift emissions are early in the longitude profile, none occurring later than 140°, while negative drifts occur late in the profile, none earlier than 190°.

To display the spectra systematically, we prepared a montage including observations from January 28, 1960 through June 28, 1960 (FIGURE 6). We eliminated the effects of the interferometer fringe patterns where they cut across the dynamic spectra, and plotted equal-intensity contours in these 28 cases. The montage represents each spectrum in its correct position on the longitude profile.

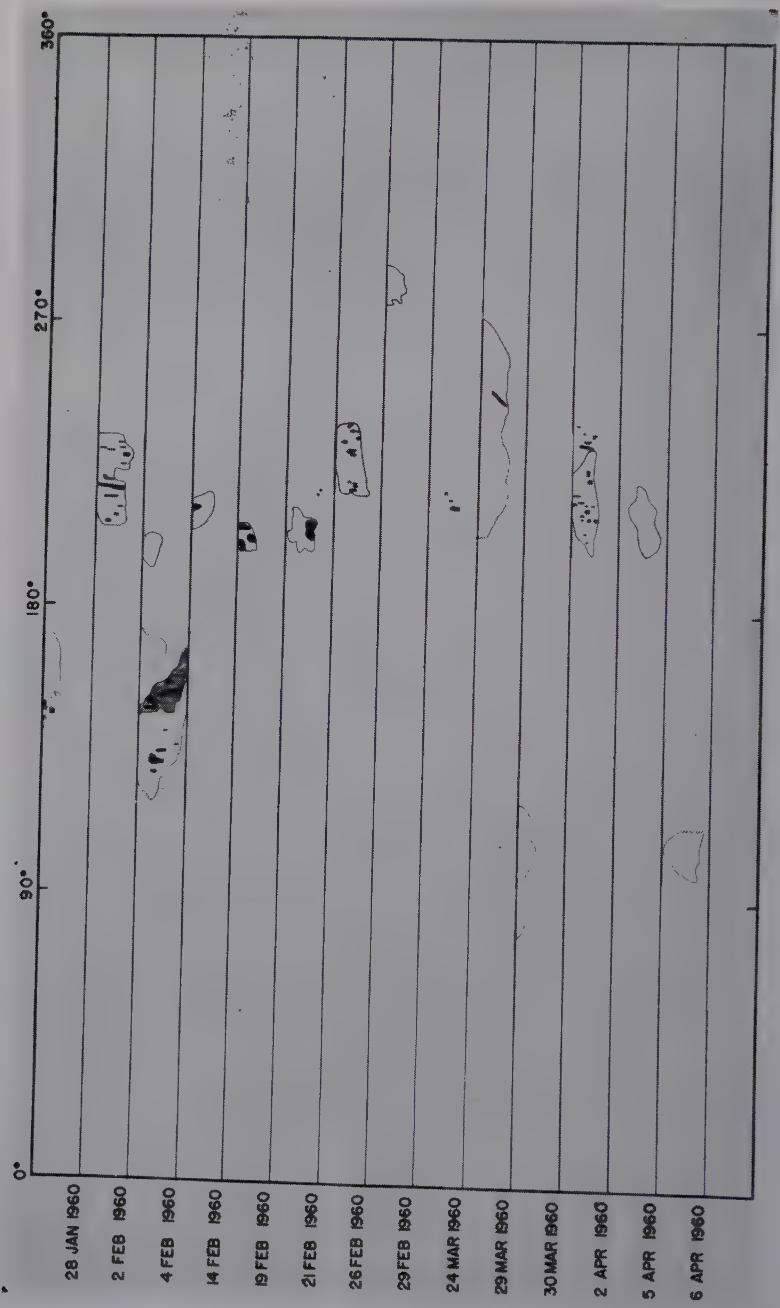
The spectral observations will reflect any effects of ionospheric scintillation just as fixed-frequency observations. One way to decide how important scintillations may be is to make simultaneous observations from widely-separated

TABLE 1

(mc./s.) min.$^{-1}$

Date	Drift	Drift rate (mc./s.) min.$^{-1}$	Longitude	Doubled structure?
February 4, 1960	+	0.5	140°	Possibly
February 26, 1960	−	0.1	220°	
March 29, 1960	−	1.0	250°	
April 5, 1960	−	0.2	190°	
May 8, 1960	+	0.7	80°	Yes. Possibly third also
May 11, 1960	−	0.7	225°	Yes
May 14, 1960	−	0.4	220°	
May 15, 1960	+	0.4	110°	Yes
May 21, 1960	−	0.2	220°	
May 22, 1960	+	0.6	75°	

receiving sites. Gardner and Shain (1958), at stations 25 km. apart, found that huge effects were the rule, although some of the bursts appeared to correlate well at the 2 stations. More recently, A. G. Smith *et al.* (1960) repeated this experiment on a baseline of one earth's radius, with essentially the same result. The results are not completely conclusive, inasmuch as ionospheric fluctuations might be expected occasionally to coincide by chance, the actual probability of coincidence depending on the statistical nature of the ionospheric irregularities. Simultaneous observations made of radio stars can support the spaced receiver technique by showing that scintillations are, in fact, present on any given occasion. FIGURE 7 displays such a comparison, made in Boulder on June 15, 1960 (local time, 105° W). The radio source Cygnus A lay in the zenith on this occasion, while Jupiter was on the meridian, about 27 degrees above our southern horizon. Cygnus A showed violent scintillations, with total durations less than 6 sec., and amplitudes exceeding 4 times the mean amplitude of the source. Our record of these scintillations is certainly limited by the integration time constant of the receiver, about three seconds. We have succeeded in observing such scintillations with the spectrograph on some


0°
90°
180°
270°
360°
28 JAN 1960
2 FEB 1960
4 FEB 1960
14 FEB 1960
19 FEB 1960
21 FEB 1960
26 FEB 1960
29 FEB 1960
24 MAR 1960
29 MAR 1960
30 MAR 1960
2 APR 1960
5 APR 1960
6 APR 1960

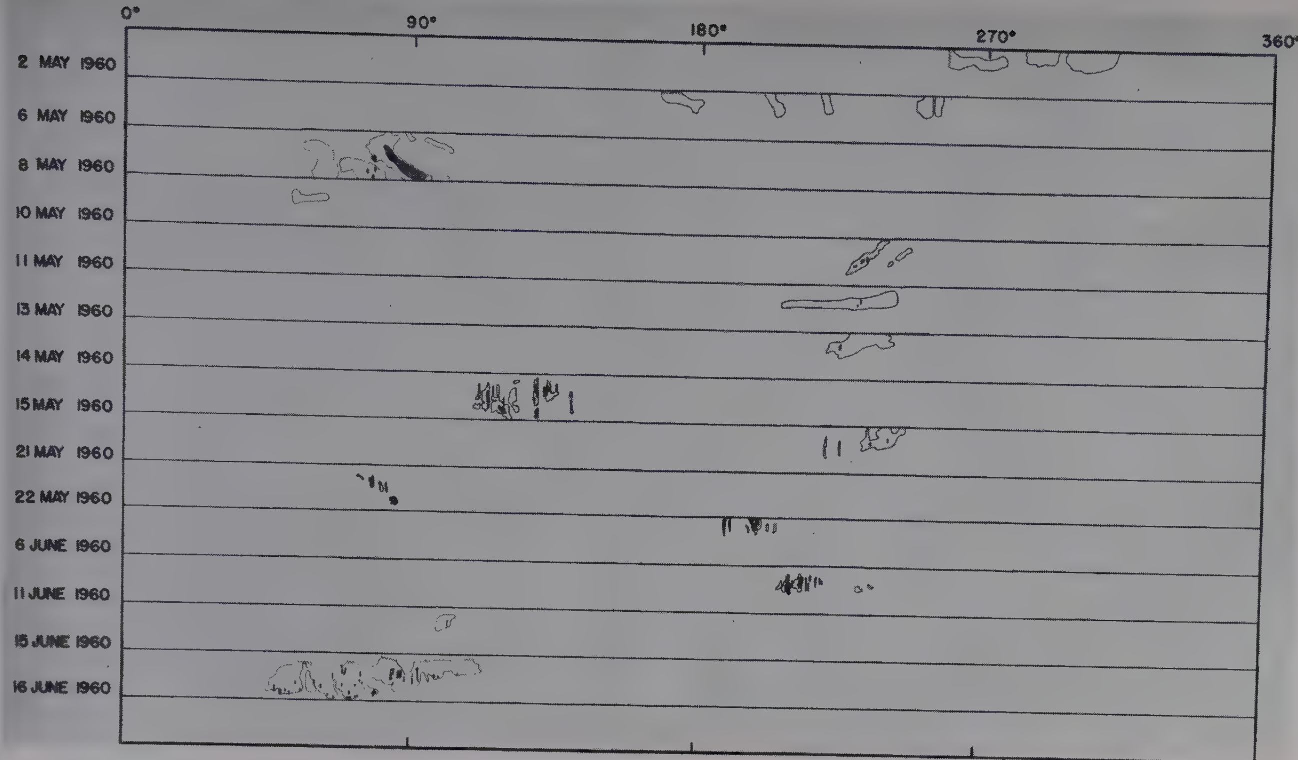


FIGURE 6. Boulder dynamic spectra of Jupiter's emission in the range 15 to 34 mc./s. These are our complete data for the period January 28, 1960, through June 16, 1960. They are displayed on the basis of longitudes in modified System III, in which the period is $9^h\ 55^m\ 29^s.5$, and the longitudes agree with System II on 1.0 January, 1957. The effects of the interferometer fringes have been removed, and equal-intensity contours in 3 qualitative steps drawn by tracing from the original records. UT dates of each record are shown at the left.

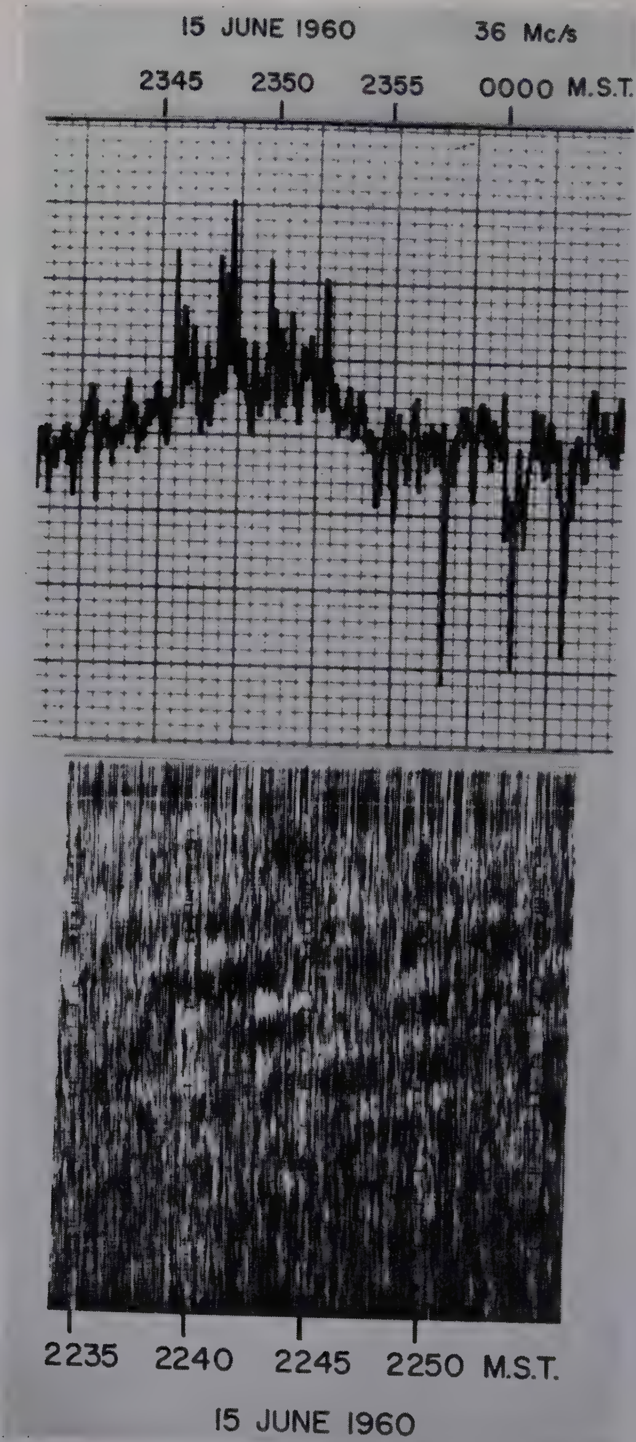

15 JUNE 1960
36 Mc/s
2345
2350
2355
0000 M.S.T.
2235
2240
2245
2250 M.S.T.
15 JUNE 1960

FIGURE 7.

occasions, and find that bandwidths may correspond to 50 per cent of the operating frequency, or more. Furthermore, as a rule, these scintillations drift.

We believe that at least on the record of June 15, 1960, the observed fluctuations in the spectrum of Jupiter's emission can be described as ionospheric scintillations, and that even reports of wideband, short-duration bursts (H. J. Smith *et al.*, 1960) may frequently, if not always, refer to the same phenomenon. It is clear, in any event, that scintillations can imitate in all essential details, including the speed, range in amplitude, and drift in frequency, the fastest observed Jupiter emissions, *excluding* the millisecond structures. At the same time, we believe that the slow-drift emissions do not correspond to any ionospheric phenomena reported by others, or observed by us. In fact, on one occasion, the event of March 29, 1960, the negative drift emission was observed simultaneously at Yale (Smith *et al.*, 1960) with about the same drift rate observed by us. In summary, we believe that at the present time the most conservative interpretation of the emission is in terms of essentially steady sources on Jupiter, which often, but not always, drift at rates of the order of one half to 1 mc./sec.·min.$^{-1}$. The sense of these drifts correlates strongly with longitude, positive drifts occurring early; negative drifts, late in the longitude profile.

Further Remarks on the Location of Jupiter's Dipole Moment

The relation between the sign of the drift and longitude suggests that a single mechanism acts at all longitudes of the decametric emission. Independent sources would be unlikely to produce symmetrically related drifts. The drift rate is very slow for time scales appropriate to Jupiter. Suppose that shock waves excite plasma oscillations, which produce radio emission as in the solar corona during Type II and Type III bursts. The time evolution presumably then results from motion of the shock through Jupiter's ionosphere. Jupiter, however, is much smaller than the sun, and its bursts drift even more slowly. An upper limit to the distance a disturbance has to travel would seem to be the thickness of the ionosphere, less than 100 km. to cross all plasma levels from zero to maximum density and back to essentially zero. The drifting emissions take 20 min. or more to cross our one octave of spectrum. The deduced velocities of the disturbance lie in the range from 10 to 100 m·sec.$^{-1}$, gentle even compared to the winds in the terrestrial ionosphere where no significant emission appears due to this cause. Thermal velocities of protons in Jupiter's atmosphere are no less than 1 km·sec.$^{-1}$. No natural source of internal excitation would seem to move slowly enough to satisfy our observations, and yet violently enough to generate shock effects producing intense radio emission.

The Čerenkov mechanism suggested above under *A Mechanism for Excitation of Jupiter's Decameter Emission* produces radiation at the gyro frequency,

FIGURE 7. A comparison of records taken almost simultaneously of Cygnus A and Jupiter, illustrating the importance of ionospheric scintillations in observations of Jupiter. The receivers were both located at the radio astronomy site of the High Altitude Observatory, near Boulder, Colo. The difference of less than 1 hour between the 2 displays is not felt to make the comparison less valid, since earlier records of Cygnus A, outside the main antenna beam, still showed strong scintillations.

varying from point to point over the surface of the planet. In the case of the earth, the field strength varies 2:1 from pole to equator, but corpuscular bombardment, except under unusual circumstances, takes place in a very restricted latitude range within the auroral zones or within the polar caps. The field strength within these zones scarcely varies at all. To produce drifting decametric emissions by corpuscular bombardment of Jupiter requires that the field strength vary along the zone. A variation of field is a rather direct consequence of an eccentric dipole moment. We already concluded on the existence of this moment from the symmetries of the decametric source; the independence of the result from the two points of view encourages us to derive further details of the moment from the decametric drifts. Strong positive and negative drifts occur roughly 180° apart in longitude. At these times, we assume that the dipole is near the limb of the planet. Longitudes near 200° seem to show no systematic drifts, and are the position of maximum rate of emission. The decentered moment lies under this longitude.

Our knowledge of the position of even the earth's auroral zone is hardly complete, but several theories (Martyn, 1951; Parker, 1958; and Warwick, 1959) using quite different approaches contain predictions of the correct latitude of the terrestrial zone. Obviously a comprehensive theory of auroras is not needed to predict which line of force defines the average position of the zone. The same formulas may then apply to Jupiter, where we know the magnetic moment and its direction in space. The unperturbed line of force to the auroral zone points at the sun in the interaction zone between interplanetary gas and the planetary magnetic field. The field strength of Jupiter's dipole is:

$$H = \frac{M_{♃}}{r^3} (1 + 3 \cos^2 \vartheta)^{1/2} \tag{4}$$

where ϑ is the angle between the moment and the field point, and r is the distance of this point from the moment Where r_m is the equatorial distance from the dipole, a line of force in the field **4** is:

$$r = r_m \sin^2 \vartheta \tag{5}$$

In the interaction region, the auroral line of force is at r_1, ϑ_1, where $r_1 = R_0/\sin \vartheta_1$, (see FIGURE 8), and $\vartheta_1 = \cos^{-1} (1/\sqrt{3}) = 54^\circ.7$. The auroral line of force then is:

$$r = (R_0/\sin^3 \vartheta_1) \sin^2 \vartheta \tag{6}$$

where R_0, again, is the equatorial distance to the interaction zone. Since, by Equation **1**, $H_0 = M_{♃} / R_0^3$, we find, on dividing this relation into Equation **4**,

$$H/H_0 = (\sin^9 \vartheta_1/\sin^6 \vartheta) (1 + 3 \cos^2 \vartheta)^{1/2} \tag{7}$$

Equation **7** defines the field strength H along the auroral line of force. If the field strength at the surface of the planet determines the frequency of the decameter emission, we can derive from Equation **7** the polar angle ϑ of the point of emission from the dipole, for each moment during a drifting emission. As in the terrestrial case, we do not expect that this theory will define the variations of frequency in more than a statistical sense. Individual events may

follow different lines of force simultaneously, or occur at a number of latitudes and longitudes along the zone. Still, we shall assume that the auroral zone, when the moment is at the limb, has a field strength of 5 gauss, corresponding to 14 mc./sec., and when the moment is at central meridian of the planet, it has a field strength of 10 gauss. If we solve Equation **7** for the corresponding angles, and Equation **6** for the radii, we find:

$$\begin{array}{lll} H = 5 \text{ gauss}; & r' = 0.748\, R_{♃}\,; & \vartheta' = 15^{\circ}\!.00 \\ H = 10 \text{ gauss}; & r'' = 0.595\, R_{♃}\,; & \vartheta'' = 13^{\circ}\!.35 \end{array} \tag{8}$$

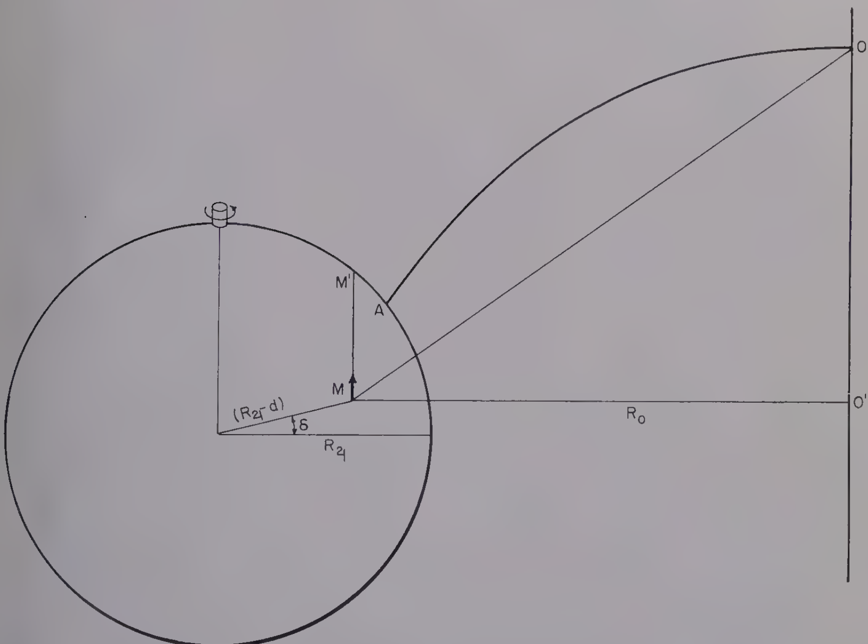


FIGURE 8. The geometrical relations between Jupiter's magnetic moment **M**, the auroral zone line of force AO, and the interaction region OO′, between solar plasma and Jupiter's magnetic field. The dip pole of Jupiter's magnetic field is at M′. The polar angle ϑ lies between the direction MM′ and the field point, and ϑ'' is the angle between MM′ and MA. The sun (or earth) lies in the direction MO′.

Introduce a coordinate system (x,y,z) whose origin is the dipole moment, with z-axis parallel to Jupiter's axis (essentially perpendicular to the ecliptic plane), and y-axis towards the earth-sun when the dipole is at central meridian (see FIGURE 9). Let d be the depth of the moment below Jupiter's surface and δ its latitude. The (spherical) surface of Jupiter takes the form:

$$R_{♃}^2 = x^2 + [y + (R_{♃} - d)\cos\delta]^2 + [z + (R_{♃} - d)\sin\delta]^2 \tag{9}$$

We have the particular values:

$$\begin{array}{lll} x' = r'\sin\vartheta'; & y' = 0; & z' = r'\cos\vartheta'; \\ x'' = 0; & y'' = r''\sin\vartheta''; & z'' = r''\cos\vartheta'', \end{array} \tag{10}$$

which can be inserted into Equation **9** to yield:

$$(r')^2 + 2\,(R_{♃} - d)\,r' \cos\vartheta' \sin\delta = (r'')^2 + 2\,(R_{♃} - d)r'' \sin(\vartheta'' + \delta) \quad (\mathbf{11})$$

$$0 = (r')^2 - 2\,R_{♃}\,d + d^2 + 2\,(R_{♃} - d)\,r' \cos\vartheta' \sin\delta \quad (\mathbf{12})$$

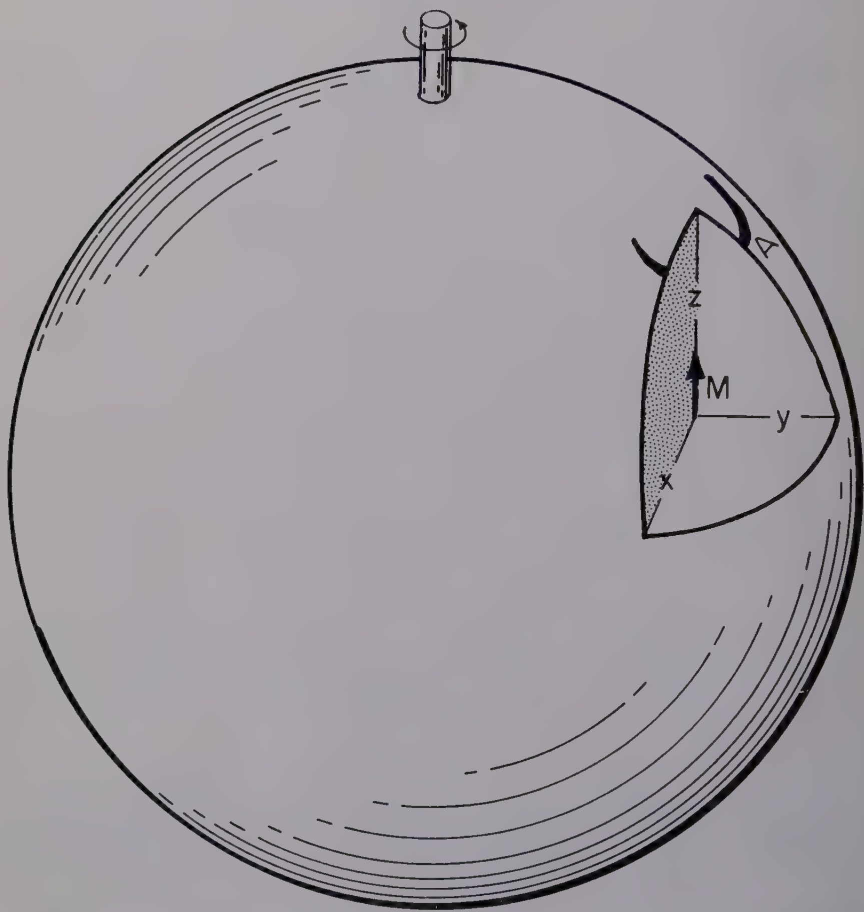


FIGURE 9. The geometry of M within the planet. The coordinate system M — x,y,z is shown in the cut-away section of the surface of the planet. Note that Mz lies parallel to the axis of rotation. My points to the sun when the dipole is on the central meridian of Jupiter. The horn-shaped auroral zone of Jupiter is shown schematically on the surface of the planet. The point A, lying on the surface, corresponds to A on FIGURE 8.

Using the values **8**, we solve Equations **11** and **12** iteratively for d and δ. There result

$$d = 0.354\,R_{♃}\,,\qquad \delta = 14^\circ.0 \quad (\mathbf{13})$$

The angle, Θ, between the line of force and the direction to the earth turns out to be $90^\circ - \vartheta - \tan^{-1}(\tfrac{1}{2}\tan\vartheta)$. Inserting ϑ' and ϑ'', we find:

$$\Theta' = 67\overset{\circ}{.}4; \Theta'' = 69\overset{\circ}{.}9 \tag{14}$$

The polarization of the emission depends on Θ according to a standard formula of magnetoionic theory (Bremmer, 1949). The axial ratio of the emergent radiation follows from Bremmer's formula for R_p, in which $x = 0$, and $y \sim 1$. With the values of Equation **14**, $R_p \sim 0.3$. About 0.1 of the total power lies in the circular mode. Orientation of the major axis of this ellipse, while it is in principle evidence for the direction of the magnetic field in Jupiter's ionosphere, will probably not prove useful in the present low-frequency case, where terrestrial Faraday effect amounts to rotations of hundreds of radians.

Summary and Conclusions

Dynamic spectra of Jupiter's emission suggest symmetry in emission drifts, around the longitude of maximum occurrence of emission at about 200°, revised System III. We suggested the presence of a single source of emission, linked to the rotation of the planet by the magnetic field. Emission takes place through an indirect excitation of gyro radiation, when high-energy electrons strike the planet's ionosphere. For simplicity's sake I have represented the magnetic field by a dipole moment, and derived the parameters of the moment and its position within the planet. The suggestion that several emission centers are present perhaps is consistent with a number of localized current sources around the planet, although this does not appear to follow necessarily from the dynamic spectra. The sense of rotation and the ellipticity of the emitted radiation are constant with an axial ratio of about 3:1.

The decimetric source rotates and should therefore produce an oscillation in the center of gravity of the emission from 500 to 3000 mc./sec. Our decametric model is completely inconsistent with Field's (1960) decimetric model requiring a very large magnetic field. Tests to discriminate between models involve precision measures of the frequency characteristics of the decimetric emission, as well as accurate positional measures.

It is gratifying to note since preparation and presentation of these notes at the URSI London Assembly (September, 1960), and of an earlier version at the AGU Washington meetings (April, 1960), that Carr *et al.* (1960) have also concluded on the existence of an eccentric magnetic field on Jupiter.

The program of Jupiter observations at the High Altitude Observatory will be continued during the apparition of 1961 with the frequency range of the spectrograph extended to cover 8 to 41 mc./sec. We hope to be able to clarify the puzzling doubling effects and say more about possible harmonics with the new data.

Summary

Dynamic spectra in the range 15 to 34 mc./sec., supported by fixed-frequency observations in the decimetric range, 400 to 3000 mc./sec., suggest a model for Jupiter's magnetic field. The model depends on a novel interpretation of the low-frequency observations: these sporadic emissions result from motion of solar and planetary particles or plasma through Jupiter's ionosphere along magnetic lines of force. The locus of the material defines Jupiter's auroral zone. The radio emission is excited very nearly at the gyro frequency of the electrons in Jupiter's magnetic field. Drifts in this frequency occur as the

rotation of the planet carries different parts of the auroral zone, with different field strengths, underneath the spatially static corpuscular stream. In support of the introduction of such an unusual, asymmetric magnetic field, which can follow only from an eccentric magnetic dipole, it has been noted that it provides a natural explanation for the asymmetric longitude profile of the decametric emission and for its consistently right-handed polarization. The deduced dipole moment has a strength of 8×10^{28} gauss·cm.^{3}, and lies parallel to the planet's rotational axis, at long. 200° (System III), lat. 14° (N or S), and 0.6 $R_{♃}$ from the center of the planet.

Acknowledgments

It is a pleasure to acknowledge discussions of these problems with J.-F. Denisse, G. B. Field, and J. A. Roberts. R. H. Lee, of the High Altitude Observatory, designed and developed the receiving equipment which made the observations possible. I am deeply indebted to him for his continuing interest in the project.

References

BREMMER, H. 1949. *Terrestrial Radio Waves*. Elsevier. New York, N. Y.
BURKE, B. F. & K. L. FRANKLIN. 1955. J. Geophys. Research. **60:** 213.
CARR, T. D., A. G. SMITH & H. BOLLHAGEN. 1960. Phys. Rev. Letters. **5:** 418.
CARR, T. D. 1959. Astron. J. **64:** 39.
DENISSE, J.-F. 1960. *Solar Radio Astronomy* (Introduction to Commission V Session on *The Sun*, 13th General Assembly of URSI, London, September. To be published by National Academy of Sciences, U. S. A.).
DRAKE, F. D. & S. HVATUM. 1959. Astron. J. **64:** 329.
DYCE, R. B. & M. P. NAKADA. 1959. J. Geophys. Research. **64:** 1163.
FIELD, G. B. 1959. J. Geophys. Research. **64:** 1169.
FIELD, G. B. 1960. Ibid. **65:** 1661.
FRANKLIN, K. L. & B. F. BURKE. 1958. J. Geophys. Research. **63:** 807.
GALLET, R. M. 1958. Publication 581, National Academy of Sciences—National Research Council (Boulder, Colorado: A Report to the National Academy of Sciences—National Research Council from the USA National Committee of the International Scientific Radio Union on the Twelfth General Assembly, August 22–September 5, 1957).
GARDNER, F. F. & C. A. SHAIN. 1958. Australian J. Phys. **11:** 55.
GINZBURG, V. L. & V. V. ZHELEZNIAKOV. 1959. Paris Symposium on Radio Astronomy. **:** 574. Stanford Univ. Press. Stanford, Calif.
KRAUS, J. D. 1958. Proc. I. R. E. **46:** 266.
LEE, R. H. 1957. Electronics. **30:** 162.
MARSHALL, L. 1956. Astrophys. J. **124:** 469.
MARTYN, D. F. 1951. Nature. **167:** 92.
PARKER, E. N. 1958. Phys. Fluids. **1:** 171.
PEEK, B. M. 1959. J. Brit. Astron. Assoc. **69:** 70.
PINES, D. & D. BOHM. 1952. Phys. Rev. **85:** 338.
RADHAKRISHNAN, V. & J. A. ROBERTS. 1960. Phys. Rev. Letters. **4:** 493.
RISHBETH, H. 1960. Australian J. Phys. **12:** 466.
SAGAN, C. & S. L. MILLER. 1960. Astron. J. **65:** 499.
SMITH, A. G., T. D. CARR, H. BOLLHAGEN, N. CHATTERTON & F. SIX. 1960. Nature. **187:** 568.
SMITH, H. J., B. M. LASKER & J. N. DOUGLAS. 1960. Astron. J. **65:** 501.
WARWICK, J. W. 1959. J. Geophys. Research. **64:** 389.
WARWICK, J. W. 1960. Science. **132:** 1250.
WARWICK, J. W. 1961. Proceedings of the Kiruna Symposium on Polar Cap Absorption, August, 1960.
WELLS, H. W. 1958. Proc. IRE. **46:** 205.
WILD, J. P. & K. V. SHERIDAN. 1958. Proc. IRE. **46:** 160.
ZHELEZNIAKOV, V. V. 1958. Russian Astron. J. **35:** 230.

THE INFLUENCE OF INFRARED ABSORPTIVE MOLECULES ON THE CLIMATE

Gilbert N. Plass
Aeronutronic, a Division of the Ford Motor Company, Newport Beach, Calif.

It is appropriate to reconsider the influence of infrared absorptive molecules upon the climate since 1961 is the centenary of the first clear statement of this theory. The physicist John Tyndall[1] wrote in 1861 that "if, as the above experiments indicated, the chief influence be exercised by the aqueous vapour, every variation of this constituent must produce a change of climate. Similar remarks would apply to the carbonic acid diffused through the air. . . . It is not, therefore, necessary to assume alterations in the density and height of the atmosphere to account for different amounts of heat being preserved to the earth at different times; a slight change in its variable constituents would suffice for this. Such changes in fact may have produced all the mutations of climate which the researches of geologists reveal. However this may be, the facts above cited remain: they constitute true causes, the *extent* alone of the operation remaining doubtful."

Tyndall might have added ozone to the list of gases that can influence the climate. The description of the physical processes of emission and absorption of infrared radiation by ozone, carbon dioxide, and water vapor is now sufficiently well understood to make it possible to say with certainty that reasonable variations in the amount of any one of these three gases in the atmosphere cause changes in the equilibrium temperature of sufficient magnitude to influence the climate.

Each one of these gases has spectral bands in the region of the infrared spectrum where an appreciable flux of energy is radiated by the earth's surface and atmosphere. When there are clear skies almost all of the downward flux of infrared radiation in the atmosphere arises from emission by the molecules of these three gases. When the concentration of any one of these gases increases, the downward radiation increases which causes the surface temperature to rise. This is popularly known as the greenhouse effect since the atmospheric molecules react with the radiation in the same manner as the glass in a greenhouse. Just as the temperature inside a greenhouse is warmer than outside, so the surface temperature of the earth is higher because of the presence of these three gases in our atmosphere than it would be without them.

The amount of these gases in the atmosphere has undergone sufficiently large variations to have had an influence upon the climate throughout the earth's history. The amount of water vapor in the atmosphere fluctuates continually at a given location as a result of the atmospheric circulation and of temperature changes. The maximum amount of water vapor that can be held by a given parcel of air decreases rapidly with the temperature. Thus a change in the average temperature at a particular location causes a relatively large change in the average atmospheric water vapor amount. The total amount of ozone and its distribution in space fluctuate continuously over a certain range of values. These changes are caused by variations in the ultra-

violet light from the sun and the stratospheric circulation. The atmospheric carbon dioxide amount is more nearly constant over short time intervals, although variations as large as 50 per cent may occur depending on whether the air mass has most recently been over an ocean, a forest, or a city. There is a continuous exchange of carbon dioxide between the atmosphere, biosphere, and the oceans. Carbon dioxide is taken from the atmosphere by the weathering of rocks, which forms carbonates, and is released from the interior of the earth by volcanoes and hot springs. When any of these factors vary, the atmospheric carbon dioxide amount also changes. The contribution of these various factors to the carbon dioxide balance has varied widely during the geological history of the earth. Thus there can be no doubt that there have been corresponding variations of the atmospheric carbon dioxide amount in past epochs.

The climate is obviously determined by a combination of many different factors. A very large number of theories of climatic change have been proposed. Most of these have attempted to explain climatic change in terms of the variation of a single variable. Rather it seems likely that first one and then another factor or combination of factors has had the predominant influence. Our task is to determine which factors were dominant during a particular geological epoch. This can be done by determining which variables influencing the climate have changed during this epoch and by attempting to make quantitative estimates of the influence of these variations when this is possible.

The influence of infrared absorptive molecules is investigated in this spirit. Certain variations in the atmospheric constituents are correlated with past geological events. The predicted influence of these events are then compared with the geological record. Quantitative estimates of the temperature change accompanying some of these variations are attempted. When the observed climatic variations agree with the reasonably simple and natural predictions of this theory, then it is reasonable to conclude that these factors may have been the dominant ones during that particular epoch. At other times completely different factors either determined the climate or had an important influence upon it. Any complete theory of climatic change must take account of the variations of many quite different factors.

First let us consider the effect of atmospheric ozone. Measurements of the ozone amount and distribution have shown that there is a considerable variation with time and place.[2] The total ozone amount in a vertical air column is usually between 0.15 and 0.40 cm. at STP. The maximum ozone concentration (as measured in centimeters of ozone at STP per km.) may occur anywhere between 10 and 30 km. The maximum value is usually about 10^{-2} cm. of ozone per km. The ozone in the lower stratosphere is not in photochemical equilibrium. The amount present is largely determined by the circulation in the stratosphere, which brings new supplies of ozone down from higher altitudes. This ozone is not destroyed rapidly since very little of the sun's ultraviolet radiation can penetrate this far down.

On the other hand, in the upper stratosphere the ozone is in photochemical equilibrium. Here the amount of ozone present is determined by the ultraviolet flux from the sun. The ultraviolet radiation in the Herzberg and Schu-

mann-Runge bands causes the ozone to be formed by first dissociating the oxygen molecules. Longer wave-length ultraviolet in the Hartley band around 2500 Å dissociates the ozone molecule. The equilibrium amount of ozone is proportional to the square root of the ratio of the intensity of sunlight in these 2 frequency intervals. Thus changes in the spectral emission of the sun necessarily cause changes in the equilibrium amount of ozone.

It has been known for many years that the ozone amount influences the infrared flux in the atmosphere. Recent discussions of the greenhouse effect due to ozone have been given by Willet,[3] Plass,[4] and Kraus.[5] A quantitative investigation[4] of the effect of ozone variations on the infrared flux shows that the downward flux at the surface of the earth, due to emission by the ozone at a wave length near 9.6 μ, increases by 0.0068 cal. cm.$^{-2}$ min.$^{-1}$ when the ozone distribution shifts from one with a maximum at 28 km. to one with a maximum at 10 km. The total ozone amounts in these distributions are 0.213 and 0.267 cm. respectively. Both of these distributions correspond closely to actual observations at a particular time and place. The corresponding change in the equilibrium temperature at the surface of the earth is 2.1° C. when there are clear skies. Probably more than half of this temperature change is due to the difference in the altitude of the maximum ozone amount, and somewhat less than half is due to the difference in total ozone amount. Because of the complexity of these infrared calculations, they have been made for only a few cases. However, it may be presumed that even larger temperature changes could be obtained by varying the total ozone amount over the observed range from 0.15 to 0.40 cm.

The atmospheric ozone provides a possible link for connecting the fluctuations in the ultraviolet light from the sun with the surface temperature of the earth. Although the ultraviolet light cannot reach the surface, it still may influence the energy balance through the infrared absorption bands of ozone. The equilibrium temperature change at the surface of the earth calculated for ozone alone is sufficiently large to cause appreciable climatic effects by itself. This value is probably increased still further by the reinforcing action of water vapor as is discussed in more detail in a later section.

The ozone can also cause radiative effects in the upper atmosphere. High ozone amounts reduce infrared cooling in the upper troposphere, which may have especial significance at cloud tops. Kraus[5] believes that this cooling would cause a lower equilibrium height for the tropical tropopause. As he has pointed out, this would tend to decrease the extent and intensity of convective disturbances and slow down the direct meridional circulation. Radiation from the sun following a solar flare has been observed[6] to cause stratospheric warming of 40° C. at the 25 mb. level. This additional energy is largely radiated away by the ozone. Calculations for a somewhat greater temperature rise at higher altitudes show that the cooling rate due to ozone increases tenfold[4] in this particular case.

There is now considerable evidence to show that there are variations in the spectral ultraviolet output of the sun. Whether there might be fluctuations over relatively long time intervals is not known. In any case the shorter term fluctuations over periods of the order of months or years do change the equilibrium ozone amount. This in turn modifies the infrared radiation in

the atmosphere. This modification is sufficient to change the equilibrium surface temperature by several degrees centigrade and to influence the atmospheric circulation and the processes of cloud formation. In brief, since ozone controls the infrared flux in the spectral region near 9.6 μ, the fluctuations in the ozone amount due to the variations of the solar ultraviolet flux and other causes has a measurable effect upon the climate.

Approximately 0.033 per cent by volume of the atmosphere is composed of carbon dioxide. There is evidence that this average amount has increased from about 0.028 per cent at the end of the 19th century to the present value.[7] This increase is probably caused by the burning of fossil fuels and the clearing of forests for agricultural use. The atmosphere of large cities often has higher than normal amounts of carbon dioxide. The carbon dioxide content of an air mass varies with its past history, in particular on whether it has been over a warm or cold ocean, over a forest or a desert, and on whether it is day or night. However, fluctuations from these causes are relatively small, and the carbon dioxide is nearly uniformly mixed in the atmosphere even at great altitudes.

Carbon dioxide interacts with the atmospheric infrared flux largely in the spectral interval from 12 to 18 μ. As the amount of carbon dioxide in the atmosphere increases, the absorption and reradiation from the carbon dioxide molecules also increases. This has the effect of increasing the downward flux of radiation, which in turn increases the temperature at the earth's surface. The change in the equilibrium temperature for a given change in carbon dioxide concentration can be calculated with some accuracy by the modern theory of band structure. Plass[8] has calculated that the equilibrium surface temperature decreases 3.8° C. when the carbon dioxide amount is halved. Kaplan[9] has calculated a value by an entirely different method which is approximately 3.0° C. for the same temperature distribution as was used by Plass.[10] These temperature changes apply for clear sky conditions. The actual temperature changes for the earth may even be somewhat larger than these values because of the action of the water vapor in reinforcing temperature changes caused by other factors. Even if the actual temperature change were somewhat smaller than the above numbers, the fluctuations in the atmospheric carbon dioxide amount has been sufficiently great over the last several billion years of geological history to have had an appreciable influence on the climate during certain epochs.

The possible variations of the many factors that determine the atmospheric carbon dioxide content have been discussed by many authors.[11-17] Let us consider how some of these factors may have varied in the past and their possible influence upon the climate.

In recent years the burning of fossil fuels has added 10^{10} tons per year of carbon dioxide to the atmosphere. It is predicted that this figure will increase to 5×10^{10} tons per year by the year 2000. In addition to burning fossil fuels, man's activities are adding additional quantities of carbon dioxide to the atmosphere by the clearance of forests, the drainage and cultivation of lands, and industrial processes such as lime burning and fermentation. If all of this carbon dioxide remains in the atmosphere, the concentration is increasing at the present time at a rate of 30 per cent a century. This would cause a

temperature rise of 1.1° C. per century, which is very close to the average temperature rise observed so far in the 20th century. However, some of this additional carbon dioxide is absorbed by the biosphere and by the oceans. The amount lost to the biosphere is not large, since the rates of decay and respiration soon increase to balance the increased rate of photosynthesis.

The carbon dioxide absorbed by the oceans is more significant. There has been considerable controversy in recent years about the length of time it takes for the oceans to come into equilibrium with the atmospheric carbon dioxide amount after a change in the latter factor. Without entering into the details of this discussion let us assume that the oceans and the atmosphere are always in equilibrium. The result of this assumption is to *underestimate* the atmospheric carbon dioxide amount from the burning of fossil fuels.

The equilibrium between gaseous carbon dioxide and the carbonates in sea water has been studied extensively and can be calculated with some accuracy.[18] The dissociation constants of sea water change with the temperature. Plass,[11] in calculating the equilibrium amounts, assumed that the average temperature of the ocean changes with the mean surface temperature on land. Curves are given in this reference for the atmospheric carbon dioxide amount as a function of the total amount of carbon dioxide or equivalent in the atmosphere-ocean system. These curves apply for various amounts of water in the ocean and are calculated both when the oceans are and are not in equilibrium with $CaCO_3$. The oceanic volume is appreciably reduced during a period of glaciation, and this factor must be taken into account in such a calculation. Over long time intervals the oceans must also come into equilibrium with $CaCO_3$. If there is an excess amount, it precipitates. If there is too little, it dissolves and, at the same time, accumulates from the rivers that flow into the oceans until its solubility product is reached. The time required for $CaCO_3$ equilibrium to occur is not known.

Let us return to the question of the burning of fossil fuels. The known reserves of coal and oil will produce 4×10^{13} tons of carbon dioxide when they are burned. A conservative extrapolation of the rate of our industrial growth shows that the known carbon dioxide reserves will have been used up in a period of the order of 500 years. The total carbon dioxide in the atmosphere-ocean system will increase from 1.32×10^{14} tons to 1.72×10^{14} tons. Even if the atmosphere-ocean system will have come to equilibrium at this time, the atmospheric carbon dioxide amount will be 0.3 per cent or 10 times its present value. The corresponding temperature rise is 12.2° C. Gradually the atmospheric carbon dioxide amount will decrease as the oceans come to equilibrium with $CaCO_3$. This presumably will take thousands of years. After this occurs the atmospheric carbon dioxide amount will be reduced to 0.11 per cent or nearly 4 times its present value. The average temperature will then be 7.0° C. warmer than today. Since equilibrium cannot be maintained between the atmosphere and oceans as additional carbon dioxide is constantly added to the atmosphere, the amount of atmospheric carbon dioxide will actually be somewhat greater than the calculated equilibrium amounts. Thus the actual temperature rise will be greater than the values calculated under the assumption of complete equilibrium. It is not yet clear whether the burning of fossil fuels or some entirely different factor is responsible for the tem-

perature rise that has occurred in this century. However, it seems certain that the additional carbon dioxide from the burning of fossil fuels will appreciably change the climate over the next few hundred years as the atmospheric amount increases.

It is interesting to attempt to find connections between the various climates that have occurred on our planet and the probable variations in the atmospheric carbon dioxide amount. There is considerable geological evidence that extensive outbursts of mountain building preceded each of the last two major glacial epochs by at least several million years. The carbon dioxide theory provides an explanation for this time lag. At the onset of a period of mountain building it is probable that increased amounts of carbon dioxide are released from the interior of the earth by new volcanoes and hot springs. The warming effect of this extra carbon dioxide offsets the cooling effect of the volcanic dust in the atmosphere and of the increase in the height of the mountains, as shown by the fact that extensive glaciers did not form during this first stage of mountain building. In fact there have been periods of mountain building during the earth's history which have not been followed by any extensive glaciation. This could be explained by assuming that the carbon dioxide released from the interior of the earth was sufficient to prevent the formation of glaciers.

When glaciers do follow periods of mountain building, the carbon dioxide theory requires a time lag of several million years before the extensive ice sheets form. During this period the carbon dioxide is being removed from the atmosphere by the process of weathering. The most active zone for the decomposition of rock occurs between the surface and the level of the permanent underground water. In mountainous country this level is farther below the surface than in flat country. Thus in mountainous country there is a considerably larger volume in which active weathering of the rocks takes place. The dominant reaction for the weathering of igneous rocks involves the formation of carbonates, thus removing carbon dioxide from the atmosphere. After a period of mountain building this increased rock weathering, acting over a period of millions of years, sufficiently depletes the atmosphere of carbon dioxide so that a period of glaciation begins. Thus the carbon dioxide theory clearly predicts an appreciable time lag between a period of mountain building and the following glacial epoch.

If the carbon dioxide amount is suddenly increased in one hemisphere, it will be distributed uniformly over both hemispheres in a relatively short time, probably less than a few decades. Thus any large scale climatic variations due to carbon dioxide must occur simultaneously in both hemispheres. The geological evidence supported by radiocarbon dating supports this view in the few cases that have been investigated. Furthermore, it is known that in the present century glaciers in both hemispheres have been receding. However, it is possible for glaciation to occur in one hemisphere, but not in the other if the topography in the two hemispheres is quite different. There might be large mountainous regions in one hemisphere where glaciers can form. If the other hemisphere had no comparable mountain ranges, the surface temperature might not be low enough to form large ice sheets.

Another factor in the carbon dioxide balance that has varied widely during the geological history of the earth is the amount of organic material trapped in

new coal and oil deposits and other organic sediments. During the Carboniferous period the amount thus trapped must have been much larger than at the present time. Most of our coal deposits were formed then. The proper conditions for the formation of these deposits were provided by the relatively flat land and the large number of marshes. After a relatively long period of time this accumulation of organic deposits appreciably reduced the amount of carbon dioxide available to the atmosphere-ocean system. It is perhaps significant that the glaciation at the end of the Carboniferous may have been the most severe in the earth's history.

Four distinct periods of glaciation during the last glacial epoch have been known to geologists for many years. Recent work has shown that there have been a dozen or so minima in the oceanic temperature in the last 600,000 years. A characteristic property of a glacial epoch is a constantly changing climate.

The carbon dioxide theory predicts such a climate: one that is undergoing constant change. The essential nonlinear element that determines the period of the oscillations and prevents the system from approaching equilibrium is the large value of the heat of fusion of water. The extra heat energy required to melt the ice sheets keeps the temperature for many thousands of years below the average value appropriate for the amount of carbon dioxide in the atmosphere.

In order to understand the causes of these fluctuations consider the sequence of events when the total carbon dioxide amount available to the atmosphere-ocean system decreases slightly, perhaps from an increased rate of rock weathering following a period of volcanism. If the reduced atmospheric carbon dioxide amount lowers the average surface temperature sufficiently, glaciers start to form. After thousands of years these glaciers grow into continental ice sheets. A considerable amount of heat energy is released to the atmosphere by the freezing of the water; this energy is eventually lost to space by radiation from atmospheric molecules. Once they have reached a certain critical size, the ice sheets seem to grow more readily, probably because of the increased albedo of the surface. The volume of water held in the ice sheets may be from 5 to 10 per cent of the volume of the oceans. However, since the reservoir of carbon dioxide in the oceans is many times the amount in the atmosphere, even this relatively small change in the oceanic volume has a large influence on the atmospheric amount. Since the ice sheets can hold only a very small amount of carbonates compared to the same volume of ocean water, the remaining water in the oceans must release carbon dioxide to the atmosphere in order to return to equilibrium. It probably takes an average parcel of ocean water several thousand years to make a complete circuit from the top to the bottom of the oceans and back again. Thus there is a time lag of at least this amount before the atmospheric carbon dioxide increases to a new equilibrium value appropriate to the smaller volume of water in the oceans.

As the atmospheric carbon dioxide increases, the average surface temperature rises. The ice sheets begin to melt. However the very considerable energy required to melt the ice sheets must be supplied by the increased downward flux of radiation from the larger number of atmospheric carbon dioxide molecules. In a typical case, where the carbon dioxide amount increases from approximately one half of its present value to its present value, the increase

in the downward flux is approximately 0.01 cal./cm.^{2} min. If one tenth of this energy is absorbed by the ice sheet and causes melting, it takes 15,000 years to melt an ice sheet 1 km. in thickness. During this period the average temperature near the ice sheets does not increase appreciably, since the energy is being used to melt the ice. The vast quantity of cold water from the ice sheets keeps the oceanic temperature low for a considerable period of time. Plastic ice flow and crustal warping may also have an influence during the melting of the ice sheets, as described by Emiliani and Geiss.[19] Plastic ice flow increases the area of the glaciers without increasing their volume. The result is to increase further the albedo. The weight of the ice sheets causes crustal warping, which decreases the average height of the land. This tends to prevent the reformation of the ice sheets when they are melting. Possibly the most important of these various time lag mechanisms is the energy absorption required for melting.

Finally, when a major fraction of the ice sheets have melted, the temperature of the oceans begins to rise. The increased oceanic volume no longer holds enough carbonates to be in equilibrium with the atmosphere. Slowly the oceans withdraw carbon dioxide from the atmosphere. As the atmospheric carbon dioxide amount decreases, the temperature falls again and the cycle starts over.

A more detailed description of the glacial oscillations together with numerical examples are given by Plass.[11] Recent measurements of the turnover times of the oceans indicate that this may not be as important a factor in the time delay as was believed originally. It is the nonlinear effects resulting from the formation of the ice sheets that are essential for the maintenance and amplification of oscillations of this sort. The time required to form a large ice sheet and then to melt it again provides an important time delay. When some other factor raises the total carbon dioxide amount to such an extent that glaciers cannot form appreciably at the beginning of one of these cycles, these oscillations can no longer occur, and the glacial epoch is at an end.

Many authors have emphasized that not only lower temperatures but also increased or at least undiminished rainfall is necessary for the formation of extensive ice sheets. It seems reasonable to expect increased precipitation when the atmospheric carbon dioxide amount decreases. Recent radiation measurements[20] have demonstrated the importance of radiational cooling at cloud tops and near the tropopause. They have indicated that atmospheric instability at cloud tops or above moist layers can be caused by radiational cooling. Calculations by Plass[8] have shown that the average temperature is lowered 2.2° C. and 1.3° C. for the upper surface of a cloud at 4 km. and 9 km. altitude respectively when the atmospheric carbon dioxide amount is halved. When the atmospheric carbon dioxide amount is decreased, the cloud top can cool more rapidly by increased radiation to space. This in turn increases the convection inside the cloud and allows it to grow in the vertical direction The time is hastened when the cloud becomes large enough and has sufficiently strong convective currents to form precipitation. In addition, the meridional circulation is increased so that the atmosphere absorbs more water vapor from the oceans. The increased cloud cover reflects more of the sun's radiations back to space so that the temperature falls still further. Thus

the interaction between the cloud cover and the atmospheric carbon dioxide amount reinforces temperature changes caused by the direct influence of the carbon dioxide on the infrared flux. When the infrared absorptive gases are the dominant factor in determining the climate, cold and wet climates tend to occur together as do warm and dry climates.

Finally, let us consider the effect of the water vapor in the atmosphere upon the climate. The water vapor is largely concentrated near the ground and decreases rapidly in relative importance with height. This occurs largely because of the decrease of temperature with height. The maximum amount of water vapor that the atmosphere can hold decreases rapidly with temperature. The highest recorded atmospheric water vapor content is 30 gm./cu. m. The atmosphere is saturated with this amount of water vapor at 30° C. On the other hand, at −50° C. the atmosphere can hold only 0.04 gm./cu. m. The amount of water vapor at a given location fluctuates through a range of values, but is effectively limited by the amount of saturated water vapor that can be held by the air at that temperature.

When the average temperature decreases, not only does the maximum allowable water vapor concentration in the atmosphere decrease, but the actual time average of the concentration decreases. This decreased amount of water vapor further reduces the surface temperature through the interaction of its infrared bands with the atmospheric infrared radiation. Wexler has calculated the effect of changes in the water vapor amount. He finds[21] for a particular water vapor distribution that the surface temperature is lowered 6° C. when the relative humidity changes from 100 per cent to 50 per cent.

Thus the effect of the water vapor in the atmosphere is to reinforce temperature changes. These changes may be caused by variations in the ozone and carbon dioxide amounts in the atmosphere. For example, if the ozone amount increases following a change in the ultraviolet radiation from the sun, the average temperature tends to rise because of increased downward infrared radiation that reaches the earth's surface. The warmer air holds more moisture on the average. Thus the downward component of the infrared radiation in the region of the water vapor bands increases. As a result the average surface temperature increases still more. Both the increased ozone and water vapor amounts prevent the cloud tops from radiating as effectively to space as before. As a result the clouds do not grow as rapidly and there is less precipitation. More direct sunlight reaches the earth's surface with the decreased cloud amount thus causing a further temperature increase. It is difficult to estimate the relative importance of these various effects, but it is significant that they all act in the same sense.

Even small temperature changes can be magnified appreciably by the various factors discussed here. Temperature changes caused by carbon dioxide variations are reinforced by the water vapor in exactly the same manner as for ozone. Water vapor can also reinforce temperature changes caused initially by other factors such as changes in the amount of volcanic dust in the air, in the continentality of the earth's surface, and in the elements of the earth's orbit around the sun.

Thus the three atmospheric infrared absorbing gases—carbon dioxide, ozone, and water vapor—act together to alter the climate in response to such external

stimuli as changes in the solar ultraviolet radiation and the various factors that enter into the carbon dioxide balance. There have been significant variations in the amount of these three gases in the atmosphere throughout the geological history of the earth. Furthermore, these variations have been large enough at certain times to have had an appreciable effect upon the climate.

Summary

Just a century ago Tyndall suggested that changes in the amount of CO_2 and H_2O in the atmosphere influence the climate. Ozone is the only other atmospheric constituent that can appreciably influence the infrared flux and thus the climate. There can be no doubt that considerable fluctuations in the amount of these three gases in the atmosphere have occurred throughout the geological history of the earth. The amount of CO_2 in the atmosphere depends on the exchange with the biological reservoir and with the oceans, on the amount released from the interior of the earth, from the weathering of rocks, and the burning of fossil fuels, and on the amount used in the deposition of fossil fuel beds. The average amount of H_2O in the atmosphere depends largely on the air temperature. The amount of O_3 depends on the intensity of ultraviolet radiation from the sun and the circulation in the upper atmosphere.

Reasonable variations in these three atmospheric gases caused by any of the above factors change the infrared flux sufficiently to induce significant climatic changes. A change in the solar ultraviolet radiation can increase the ozone amount in the upper atmosphere sufficiently to raise the temperature at the surface of the earth by several degrees Centigrade. Changes in the CO_2 amount can provide simple explanations of the time lag between the periods of mountain building and the onset of glaciation, the increased precipitation at the beginning of a glacial period, the simultaneous occurrence of glaciation in both hemispheres, the oscillations of the climate during a glacial epoch, and the general warming of the climate during the present century. The effect of H_2O is to reinforce the temperature changes caused by CO_2 and O_3. When the temperature rises, the greater amount of H_2O in the atmosphere increases the downward infrared flux at the earth's surface and thus increases the temperature. At the same time a decreased cloud cover due to a lower rate for radiational cooling at the cloud tops allows more direct sunlight to reach the earth's surface. The result is an additional increase in temperature. Thus the interrelationships between these atmospheric gases and the factors that control the amount of these gases in the atmosphere result in significant temperature changes which may have been caused initially by seemingly small changes in our environment.

References

1. Tyndall, J. 1861. On the absorption and radiation of heat by gases and vapours, and on the physical connection of radiation absorption and conduction. Phil. Mag. **22:** (Series 4) 169–194; 273–285.
2. Craig, R. A. 1950. The observations and photochemistry of atmospheric ozone and their meteorological significance. Meteorological Monographs, **1**(2). American Meteorological Society. Boston, Mass.
3. Willet, H. C. 1953. Climatic Change: Evidence, Causes, and Effects. H. Shapley, ed. Harvard Univ. Press. Cambridge, Mass.

4. Plass, G. N. 1956. The influence of the 9.6 micron ozone band on the atmospheric infrared cooling rate. Quart. J. Roy. Meteorol. Soc. **82:** 30–44.
5. Kraus, E. B. 1960. Synoptic and dynamic aspects of climatic change. Quart. J. Roy. Meteorol. Soc. **86:** 1–15.
6. Scherhag, R. 1960. Stratospheric temperature changes and the associated changes in pressure distribution. J. Meteorol. **17:** 575–582.
7. Callendar, G. S. 1958. On the amount of carbon dioxide in the atmosphere. Tellus. **10:** 243–248.
8. Plass, G. N. 1956. The influence of the 15 micron carbon dioxide band on the atmosphe ic infrared cooling rate. Quart. J. Roy. Meteorol. Soc. **82:** 310–324.
9. Kaplan, L. D. 1960. The influence of carbon dioxide variations on the atmospheric heat balance. Tellus. **12:** 204–208.
10. Plass, G. N. 1961. The influence of carbon dioxide variations on the atmospheric heat balance. Tellus. **13**.
11. Plass, G. N. 1956. The carbon dioxide theory of climatic change. Tellus. **8:** 140–154.
12. Arrhenius, S. 1896. On the influence of carbonic acid in the air upon the temperature of the ground. Phil. Mag. **41:** 237–276.
13. Chamberlin, T. C. 1897–1899. A group of hypotheses bearing on climatic changes. J. Geol. **5:** 653–683; The influence of great epochs of limestone formation upon the constitution of the atmosphere. J. Geol. **6:** 609–621; An attempt to frame a working hypothesis of the cause of glacial periods on an atmospheric basis. J. Geol. **7:** 545–584; 667–685; 751–787.
14. Callendar, G. S. 1938. The artificial production of carbon dioxide and its influence on temperature. Quart. J. Roy. Meteorol. Soc. **64:** 223–237.
15. Callendar, G. S. 1949. Can carbon dioxide influence climate? Weather. **4:** 310–314.
16. Plass, G. N. 1956. Effect of carbon dioxide variations on climate. Am. J. Phys. **24:** 37′–387.
17. Plass, G. N. 1959. Carbon dioxide and climate. Scientific Am. **201:** 41–47.
18. Sverdrup, H. U., M. W. Johnson & R. H. Fleming. 1942. The Oceans: Their Physics, Chemistry, and General Biology. Prentice-Hall. New York, N.Y.
19. Emiliani, C. & J. Geiss. 1957. On glaciations and their causes. Geologische Rundschau. **46:** 576–601.
20. Kuhn, P. M. & V. E. Suomi. 1960. Infrared radiometer soundings on a synoptic scale. J. Geophys. Research. **65:** 3669–3677.
21. Wexler, H. 1953. Climatic Change: Evidence, Causes, and Effects. H. Shapley, Ed. Harvard Univ. Press. Cambridge, Mass.

TEMPERATURE AND MAGNETIC FIELD

L. Egyed

Geophysical Institute, Eötvös University, Budapest, Hungary

The connection between solar activity and terrestrial climate is a long-discussed problem. As early as 1873 Köppen[1] sought a correlation between the annual mean equatorial temperature and the sunspot numbers, but he was unable to prove the relation for latitudes other than zero. The demonstration of a correlation between the weather in a given area and solar activity is even more difficult.

Z. Berkes[2] has studied the statistical correlation of the 27-day period of sunspot activity with wind velocities in Budapest over a 5-year period. He found that, excepting a phase shift of about 1 week, the trend of the 2 variables is the same (FIGURE 1).

On the other hand, there exists a close connection, established by J. F. Cox,[3] between the east-west component of wind velocity and the fluctuations of the angular velocity of the earth's rotation. The secular fluctuation of the angular velocity of the earth shows a trend similar to the secular variations of the terrestrial magnetic field. Therefore it seems necessary to look into the possible correlation of the local temperature changes and the changes of the magnetic parameters. One of the possibilities for checking such a correlation is to find out whether there exists a 27-day recurrence in the temperature distribution, too, and see whether it is related to the 27-day recurrence of magnetic activity.

The correlation between the sunspot numbers and the daily maximum temperature was studied by S. W. Visser,[4] who sought to demonstrate the 27-day period by harmonic analysis. However he had studied the trend and distribution of the daily temperature maximum and obtained no information as regards the nature and structure of the temperature fluctuation. Recognizing this lack, I attacked the problem in a different way. We have selected for, roughly, one sunspot period, namely 1946 through 1959, the days that were characterized by values equaling or surpassing 1.7 of the magnetic character figure C_i.

On the other hand, we have subjected to statistical analysis the soil temperature data from 50 centimeters depth at one of the Hungarian meteorological stations. The depth of 50 cm. served to eliminate the temperature fluctuations of the surface, possibly due to a variety of factors, and to yield a characteristic daily average. In order to eliminate the temperature changes due to the seasons, we have utilized, instead of the temperature data themselves, 1-day temperature differences. The statistical average was computed by taking the day with a character figure above 1.7 as zero time and by writing up the daily temperature differences $\Delta\theta$ for 10 days before and 45 days after zero day. Subsequently the values of temperature and of the magnetic character figure, belonging to days of equal serial number (that is, situated at a given span of time before or after zero day), were added. The sums thus obtained can be expressed as $m_k = \sum_i C_i^k$ and $a_k = \sum_i \Delta\theta_i^k$ respectively.

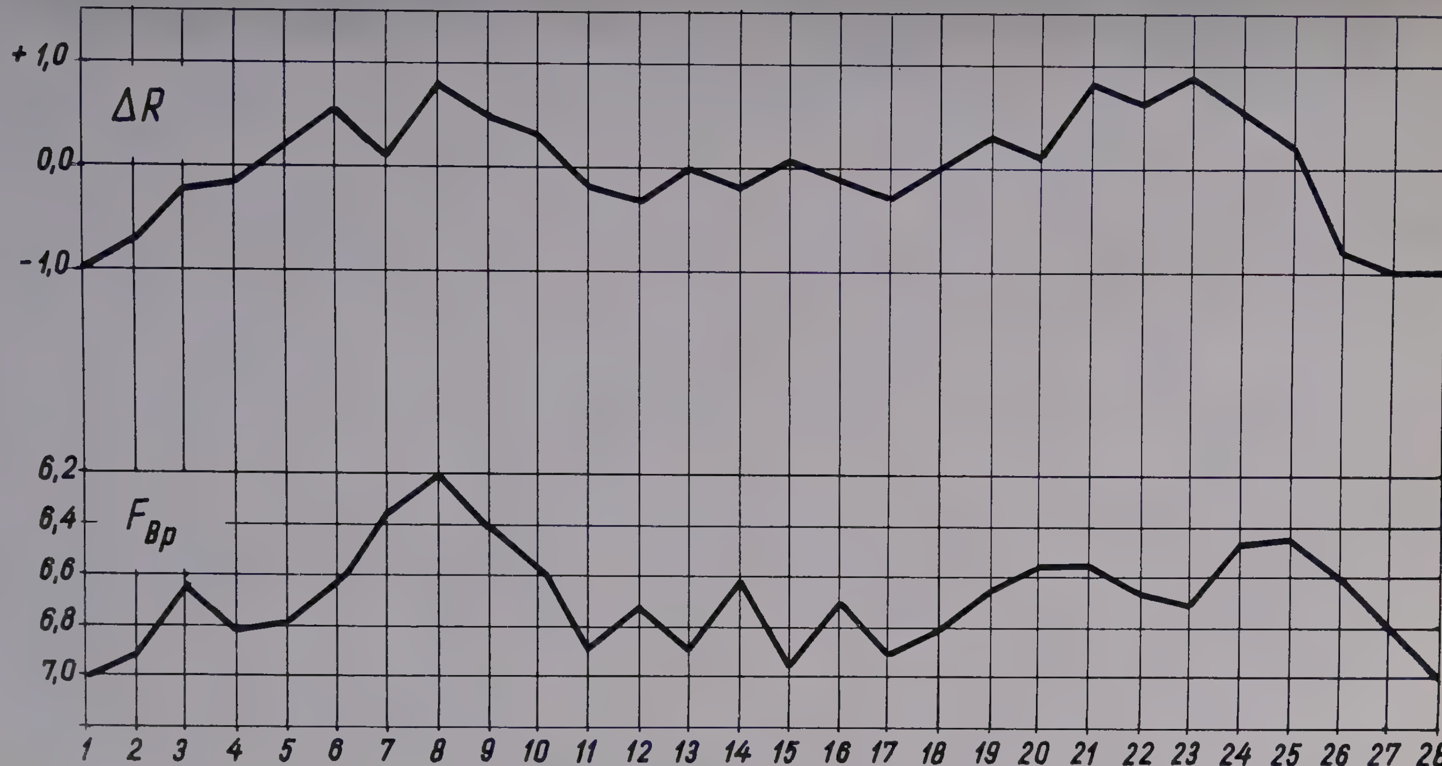
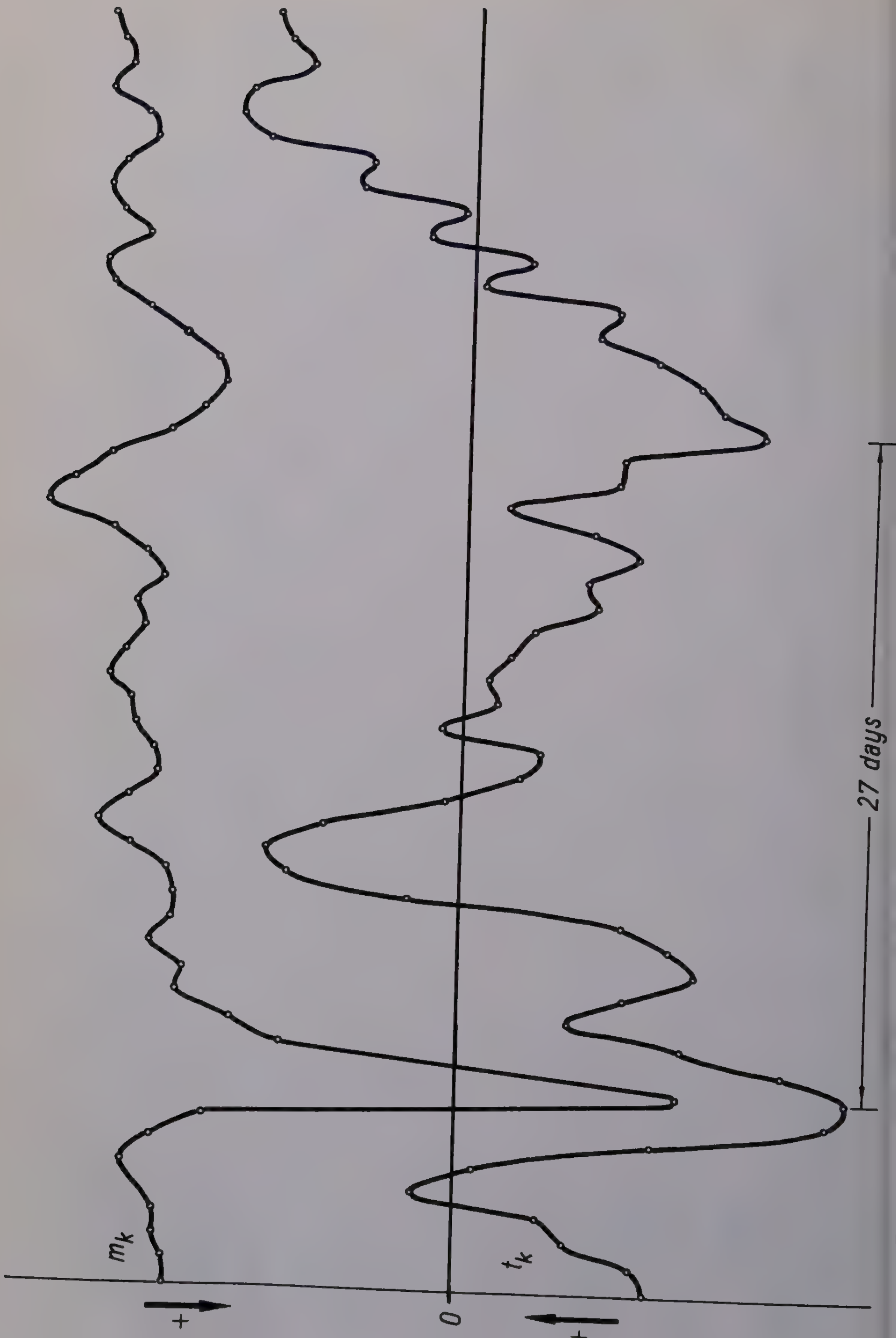


FIGURE 1. The 27-day variation of wind velocities in Budapest, Hungary, as compared with the 27-day period of solar activity.


m_k
t_k
+
0
+
27 days

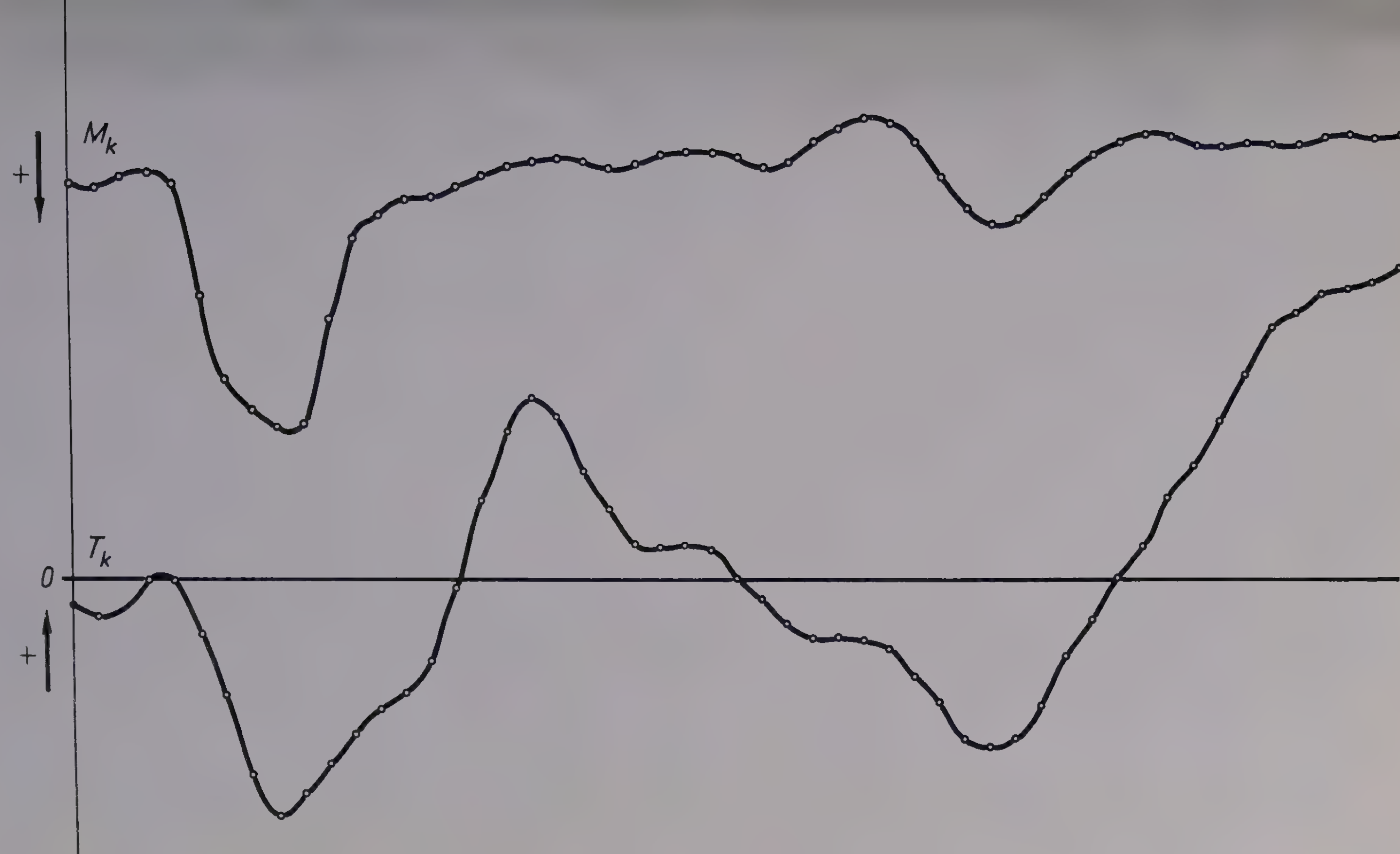


FIGURE 3. Variation of five-day running averages of soil temperature as compared with five-day running averages of the magnetic indices, C_i.

From the latter we have derived the values

$$t_k = \sum_{-8}^{k} a_k$$

of the average temperature trend curve. The results for m_k and t_k, respectively, are shown in FIGURE 2.

The t_k diagram possesses 2 characteristic minima. Between these, the time difference is an exact 27 days. The extremes of both diagrams coincide, although the second maximum of m_k is less pronounced. However in the temperature curve the second maximum corresponding to the 27-day period is much sharper than in the curve of the magnetic data, although the two curves present many similarities even as regards smaller details.

The reality of the extremes is still further enhanced by considering, instead of the a_k values, 5-day running averages, such as

$$A_k = \sum_{\lambda=-2}^{+2} a_{k+\lambda}$$

and forming the diagram of the terms

$$T_k = \sum_{-8}^{k} A_k$$

The correlation of these data with the similarly formed running averages of the magnetic C_i figures, M_k, is shown in FIGURE 3. The T_k diagram possesses two minima of a time difference of 28 days, which shows the real period to be between 27 and 28 days.

These diagrams suggest that, on the statistical average, the temperature will assume a minimum on a day of intense magnetic activity and will, thereafter, rise markedly to a maximum after the lapse of about 10 days. On the other hand, the diagram as a whole suggests an over-all rising trend. The same result is arrived at in case the temperature data are arranged according to a 27-day period independent of magnetic activity. This indicates that between 1946 and 1959 the surface temperature of the earth has undergone a steady rise.

Consider, now, that according to Holmberg,[8] the sun exerts an accelerating torque upon the earth by the mediation of the atmosphere. The intensity of this relationship must depend on insulation, and must therefore show a parallel trend to sunspot activity and to magnetic activity. The fluctuation was seen to be manifested in the wind velocities. The problem arises, then, of whether the parallel trend of the secular variation of the magnetic field and of the angular velocity of the earth, as demonstrated by Stoyko, Barta, and Vestine,[5,6,7] could not eventually, partly or entirely, be the result of these atmospheric influences. The solution of this problem would, in my opinion, also lead to a better understanding of the origin of the terrestrial magnetic field.

References

1. KÖPPEN, W. 1873. Österr. Meteorol. Z. **8**.
2. BERKES, Z. 1955. Methoden und Ergebnisse der langfristigen Wettervorhersage in Ungarn. Acta Agron. Acad. Sci. Hung. **5:** 79–98.

3. Cox, J. F. 1951. On some possibilities offered by the study at planetary scale of terrestrial phenomena. Trans. Am. Geophys. Union. **32:** 536, 537.
4. Visser, S. W. 1958. The 27-day period in United States temperatures. Trans. Am. Geophys. Union. **39:** 835–844.
5. Barta, G. 1954. A földmágneses tér évszázados változásának 44-éves periodusáról (On the 44-years' period of the geomagnetic secular variations). Geofiz. Közlemények. **3:** 3–26.
6. Vestine, E. H. 1953. On variations of the geomagnetic field, fluid motions, and the rate of the earth's rotation. J. Geophys. Research. **58:** 127–145.
7. Stoyko, N. 1951. Paris C.-R. Acad. Sci. **223:** 80; ibid. **234:** 1798–1800.
8. Holmberg, E. R. R. 1952. M.N.R.A.S. Geophysical Suppl. **6:** 325.

SUNSPOT CYCLE CORRELATIONS

David Williams
Consolidated Edison Co. of New York, Inc., New York, N.Y.

The first correlation between sunspot activity and terrestrial phenomena appears to have been made in 1801 by the elder Herschel[1] who, from his studies of 6 periods between 1650 and 1800, concluded that, in periods with little or no sunspots, wheat was scarce and hence prices were high; conversely, in periods of abundant sunspots, crops were abundant and prices were low. Herschel consequently correctly predicted that the next period of abundant sunspots would be accompanied by abundant crops. The mean relative sunspot numbers increased from a low of 4.1 during 1798 to a high of 47.5 during 1804. Agricultural production increased enormously from the low reached during the wet summer of 1799.

Although the periodicity of the sunspot cycle was not announced by Schwabe[2] until 1844, when he estimated its average length to be approximately 10 yr., the electromagnetic nature of sunspots was first suggested in 1833 by the younger Herschel.[3] However, it was not until 1908 that Hale and his associates at Mt. Wilson, Calif., were able to obtain experimental proof of the existence of a magnetic field around a sunspot (Skilling and Richardson[4]).

The Sunspot Cycle and Electromagnetic Correlations

The close correlation between sunspot activity and variations in the earth's magnetism was independently established during 1850 to 1852 by John Lamont of Munich, Germany, Edward Sabine of London, England, Alfred Gautier of Geneva, Switzerland, and Rudolph Wolf of Berne, Switzerland (See[5] and Abetti[6]). However, the first observed example of a direct influence of a solar event upon the earth was made simultaneously and independently by Carrington and Hodges during the eruption of a large group of spots between August 28 and September 3, 1859. Chapman and Bartels[7] quote Balfour Stewart of the Kew Observatory, London, England, to the effect that this eruption was followed almost simultaneously by a terrestrial magnetic disturbance, which increased into a magnetic storm of exceptional intensity, accompanied by a bright aurora borealis that was visible down to low latitudes, and a pronounced disruption of telegraphic communications.

Stetson[8] concluded from a comparison of sunspot and magnetic activity between 1850 and 1945 that "the close correlation of changes in the earth's magnetism with the coming and going of sunspots is one of the best established connections between sunspots and the earth that science knows." Nevertheless, Abetti[6] cautions, "This is clearly shown when *annual* averages are used, but becomes confused in the monthly averages and disappears in the daily."

Abetti[6] credits G. B. Donati as the first to state that auroras must be dependent on causes located in the sun, based on the latter's study of the great aurora of February 4 to 5, 1872. Since then, a very close correlation between sunspot activity and auroras has been noted. Abetti[6] also credits Guglielmo Marconi as the first to note the interference with radio transmission on Sep-

tember 20, and October 24, 1927, coincident with the appearance of large sunspots and intense auroras. Telegraph lines and oceanic submarine cables were also rendered unusable on those dates.

However, Anderson[9] states, "In 1923, in connection with the systematic measuring program of transatlantic radio transmission inaugurated by the Bell System, the association of abnormal radio transmission and disturbances in the earth's magnetic field was soon discovered and is believed to be the first direct evidence of this kind." The importance of this association prompted the United States Bureau of Standards, Washington, D.C., to establish a regular Radio Warning Service, that continued the work of the National Defense Research Committee set up in July, 1942, under the direction of D. H. Menzel and W. O. Roberts. In 1957, Fraser[10] showed that the electron density of the ionosphere waxes and wanes in step with the 11-yr. sunspot cycle.

The first notable disturbance to electric power transmission systems occurred during the great magnetic storm of Easter Sunday, March 24, 1940. Davidson[11] reported the effects on 10 power systems located in the United States in New England, New York, eastern Pennsylvania, southern and eastern Minnesota and, in Canada, in Ontario and in Quebec, P.Q. Germaine[12] reported that this storm interrupted all overseas, radiotelephone circuits, service to ships at sea, and a number of long-distance land-telephone and other communication services, such as the telephotograph network and major network broadcasting facilities.

The second disturbance occurred during the magnetic storm of Sunday, November 13, 1960, during which *Electrical World*[13] reports that the 138,000-v transmission systems of all the local New York, N.Y., utilities, as well as those in upstate New York and the Hydro-Electric Power Commission of Ontario, Canada, were affected.

The mechanism through which a solar disturbance affects large electric power systems is believed to be the following. Changes in the earth's magnetism, probably caused by the ejection of charged particles from the disturbed portion of the sun, cause differences in earth potentials. Differences of earth potential at widely separated points, where wye-connected transformers in the transmission system have their neutrals grounded, cause direct currents to flow in such a manner as partially to saturate the transformer cores. The excitation requirements of the transformers are thereby increased and the system voltage drops in an erratic pattern.

The effects of magnetic storms on underground electric cables had been given little or no consideration until August 17, 1959, when failures on 7 of 20 13,800-v cables serving the Central Park (New York, N.Y.) Network Area of the Consolidated Edison Co. of New York, Inc., necessitated the interruption of electric service to about 500,000 people. A severe magnetic storm that also blacked out radio communications was raging at the time.

Research over the period 1944 to 1960 inclusive revealed that there was a significant degree of correlation between magnetic activity, temperature, and underground cable failures on the 2300 to 138,000 v system of the Consolidated Edison Co. (FIGURE 1). A 7-day moving average of the K-Index of geomagnetic activity and of cable failures for the years 1955 to 60 showed evidence of

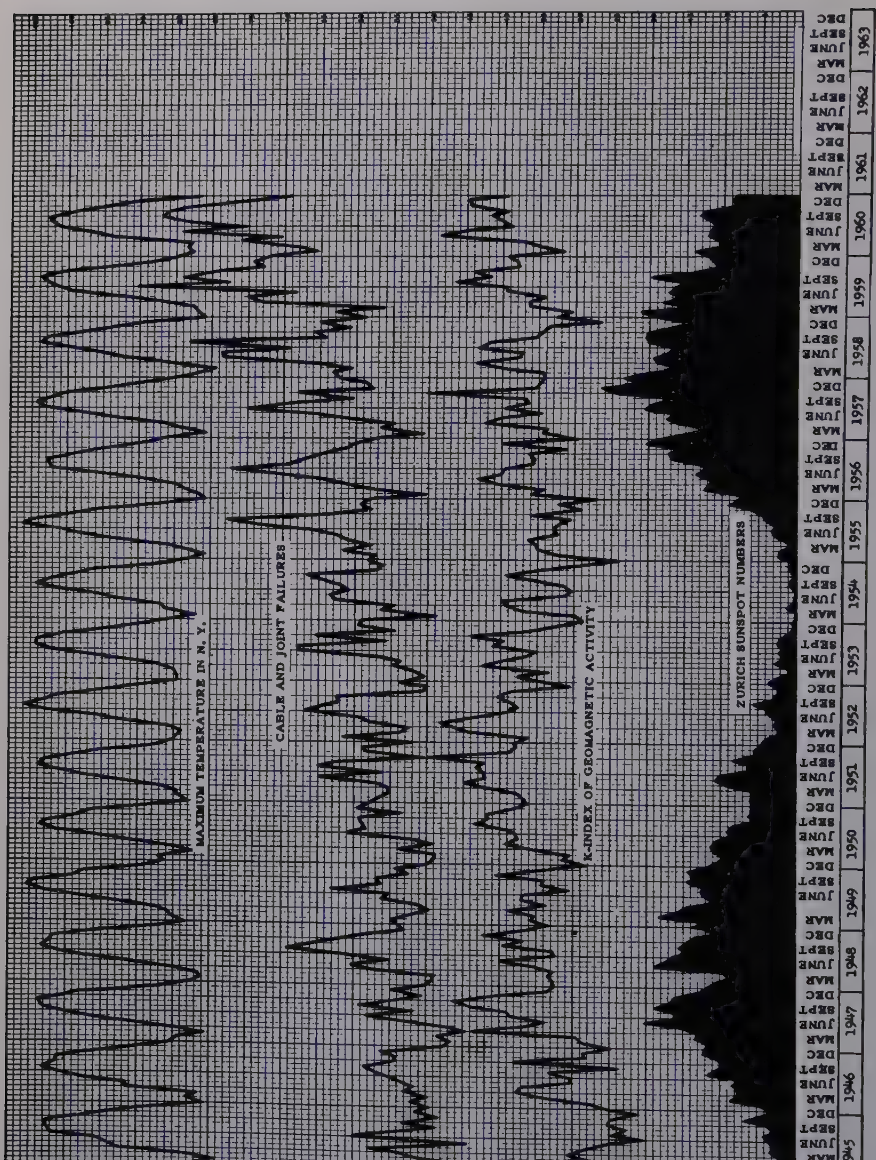


FIGURE 1. Sunspots, geomagnetic index, cable failures, and maximum temperatures

a 13.5 day cycle. Ward (F. W. Ward, Jr., this monograph) also found a 13 to 14 day cycle in a variance spectrum analysis of the K-Index of geomagnetic activity. This period is one half the 27-day solar rotation period, and indicates the possibility of tidal waves in the solar atmosphere in each solar revolution.

The coefficients of correlation for the monthly average of the daily values of cable failures, K-Index, maximum temperature in New York, N.Y., and the monthly load of the Consolidated Edison Co. are shown in TABLE 1.

Although the exact mechanism involved in the interaction of temperature, geomagnetic activity, and underground cable failure is not completely understood, the following hypothesis is offered. Experiments conducted during and since the International Geophysical Year showed that the over-all electron density of the earth's outer atmosphere (the F_2 layer) is roughly twice as great in December as in June.[14] Since the power of an ionized layer of the ionosphere to reflect radio waves back to the earth depends upon the number of free electrons present and their density, more favorable short-wave long-distance radio communication via the F_2 layer occurs in winter than in summer.

TABLE 1

COEFFICIENTS OF CORRELATION FOR CABLE FAILURES, MAGNETIC ACTIVITY, TEMPERATURE AND LOAD

(1944 to 1960, Inclusive)

	Correlation coefficient (CC)	Probable error (PE)	Ratio: CC/PE	Odds against chance relationship
Cable failures and load	0.719	±0.0223	32.2	Billions to 1
Cable failures and temperature	0.417	±0.0390	10.7	Billions to 1
Cable failures and magnetic activity	0.396	±0.0398	9.9	Billions to 1
Magnetic activity and temperature	0.169	±0.0459	3.8	98 to 1

Similarly, in the case of the oil-impregnated paper-insulated underground cables, the molecular density of the oil in the cable is greater in winter than in summer. Hence the oil keeps out any air or water that may have entered the cable through a break in the lead sheath. However, in summer, the oil expands with rising temperatures, and the molecular density of the oil falls, permitting the entrance of air and water, which deteriorates the insulation to the point where a geomagnetic disturbance precipitates electrical failures (geomagnetic disturbances have no noticeable effect on undamaged cable, hence it would seem that magnetic activity is the catalyst that weeds out damaged cable).

The Sunspot Cycle and Climatology

Herschel[1] reported in 1801 that years of remarkably abundant or deficient spots have also been remarkable respectively for their high or low general temperature and, especially, for abundant and deficient harvests. Wolf[15] reported in 1852, from an examination of the Chronicles of Zurich from 1000 to 1800 A.D., that "years rich in solar spots are in general drier and more fruitful than those of an opposite character, while the latter are wetter and more stormy than the former." He also found that during years of maximum sunspots

there have been on the average 6 to 8 violent hurricanes per year, while the average number during sunspot minimum is only 1 or 2 per year.

Pogson[16] in 1858 stated that a relationship existed between sunspot phenomena and atmospheric conditions on the earth, as evidenced by observations made in equatorial regions. In a report to the Indian Famine Commission of 1878 to 1879, Pogson traced an intimate connection between sunspot frequency, rainfall, and grain prices in Madras, India. Lockyer[17] in 1878 also traced a relationship between sunspots and rainfall in southern India. Garriott[18] in 1902 stated, "the sun's magnetic influence, stretching out and embracing the earth, varies the earth's magnetism and gives rise to weather changes."

Beveridge[19] in 1922 stated that many regular periodic movements affecting weather and crops may be accounted for through similar oscillations on the sun, moon, or even the planets. Clayton[20] in 1923 reported that in equatorial regions: (1) atmospheric pressure is less at sunspot maxima than at sunspot minima; and (2) temperatures are distinctly lower during sunspot maxima and higher at sunspot minima.

In 1933, Clough[21] stated that 11-, 37-, 83-, and 300-yr. sunspot cycles are apparent in auroral data, frequency of severe winters, frequency of Chinese earthquakes, flood and low stages of the Nile, tree growth in Arizona and California, and wheat prices in England, over a period of 1400 yr.

Clayton[22] concluded in 1943 that "in general the temperature in winter in continental interiors oscillates inversely with sunspots. In summer the relation is not well marked, but except in the eleven year period it tends to oscillate with the sunspot values. The amplitude of the oscillation in sunspots is greatest with the eleven year period, but in temperature and amplitude of oscillation is greatest with the eight year period. This fact is important because it now appears that the eleven year period in the solar constant is not so important as are the oscillations of shorter periods."

Huntington[23] states, "Cyclonic storms represent the effect of the electromagnetic field of the Sun and the solar system superposed upon the still greater effect of the Sun's heat." Abbot[24] states, "It is well known that sunspots are like machine guns, in that they bombard space, including, of course, the earth, with electric ions. This bombardment is very active at times of maximum numbers of sunspots. It is also well known that electric ions, which in our atmosphere, besides reflecting radio waves around the earth so that we get programs from great distances, in addition act as centers of condensation for the water vapors of the atmosphere and so promote cloudiness, and doubtless also rain. Clouds, of course, also alter temperatures. So in this way, the 11⅓ year sunspot cycle becomes a weather cycle."

Gillette[25] explains that weather disturbances are caused by influxes of solar electrons in five ways: (1) by generation of atmospheric currents in accordance with Faraday's principle of magnetic rotation of electric currents; (2) by the tendency of electrons to cause condensation of atmospheric moisture in accordance with Wilson's principle; (3) by increased windiness that increases oceanic evaporation and rainfall; (4) by increased evaporation due to increased electronic charge of water; and (5) by reduced influx of solar heat, due to reflection of radiant waves by atmospheric electrons and to absorption and scattering by atmospheric moisture.

Sunspots and the Business Cycle

The English economist, W. Stanley Jevons,[26] following up Herschel's[1] theory, in 1875 attempted to correlate an 11.11-yr. cycle in fluctuations in the prices of wheat, barley, oats, beans, peas, vetches, and rye with sunspot cycles of the same length. In 1878, Jevons[27] withdrew his former paper and proposed a 10.43-yr. cycle in crops as being related to 10.45-yr. sunspot cycles. His son, H. S. Jevons,[28] in 1909 modified his father's theory by stating, "It is not, as used to be supposed, the 11 year, or sunspot period which is the important factor in determining the cycle of trade and the occurrence of commercial crises. Probably the sunspot period does have some effect; but it is the 3½ year, or Solar prominence period with which we are primarily concerned in accounting for trade fluctuations."

It was not, however, until 1923 that statistical proof of the existence of a 3⅓-yr. or 40-mo. cycle in British and American indexes was established by Kitchin.[29] Thereafter, this cycle in economics was called the "Kitchin" cycle.

Economists, however, tended to belittle the Herschel-Jevons sunspot theory of the business cycle until 1934, when Garcia-Mata and Shaffner[30] reported a startlingly high degree of correlation between solar activity and total production, exclusive of agriculture, for the period from 1875 to 1930. Exceptions were found only during the depressions of 1903 to 1904 and 1913 to 1914, which were said to be due to the enormous quantity of volcanic dust blown into the atmosphere during the volcanic eruptions of Mount Pelée Martinique I., French West Indies, in 1902 to 1903, and Mount Katmai, Alaska, in 1912 to 1913.

Silberling[31] in 1943 summarized the sunspot theory of the business cycle as follows: "The reasoning, as expressed by Sir William Herschel, early in the nineteenth century, and later by W. Stanley Jevons, is that variation in the radiation or heat from the sun influences the yields of important farm crops and hence farm prices and income; these in turn affect the state of general trade and industry. Still more recent has been the attempt to read a new meaning into the roughly cyclical pattern of sunspots by suggesting that it is not the effect of sunspots, or alterations in physical radiation from the sun, upon rainfall or temperature or crop production, but rather their direct effect upon human psychology that accounts for the ups and downs in economic motivation. Briefly, the theory is that the sun transmits varying amounts of ultra-violet ray emanations, and these, in turn, have a physiological or neurological effect upon animal organisms that might conceivably be capable of bringing about alterations in moods, altitudes, and promotional energy."

Silberling[31] prepared a chart showing the relationship between the sunspot cycle and a General Business Index for the period 1750 to 1940, which the writer has extended to 1960 by adding the Cleveland Trust Co. Index. This is reproduced as FIGURE 2, and shows a significant degree of correlation. Of the 20 peaks in sunspot numbers, only 6 or 30 per cent were out of phase with the Business Index.

In 1950, I called attention to the startling fact that the United States was engaged in a war or was going through depression every time the polarity of the sunspots changed from the Southern Solar Hemisphere to the Northern Solar Hemisphere on Anderson's[32] chart of the 22¼-yr. sunspot period, which is reproduced as FIGURE 3.

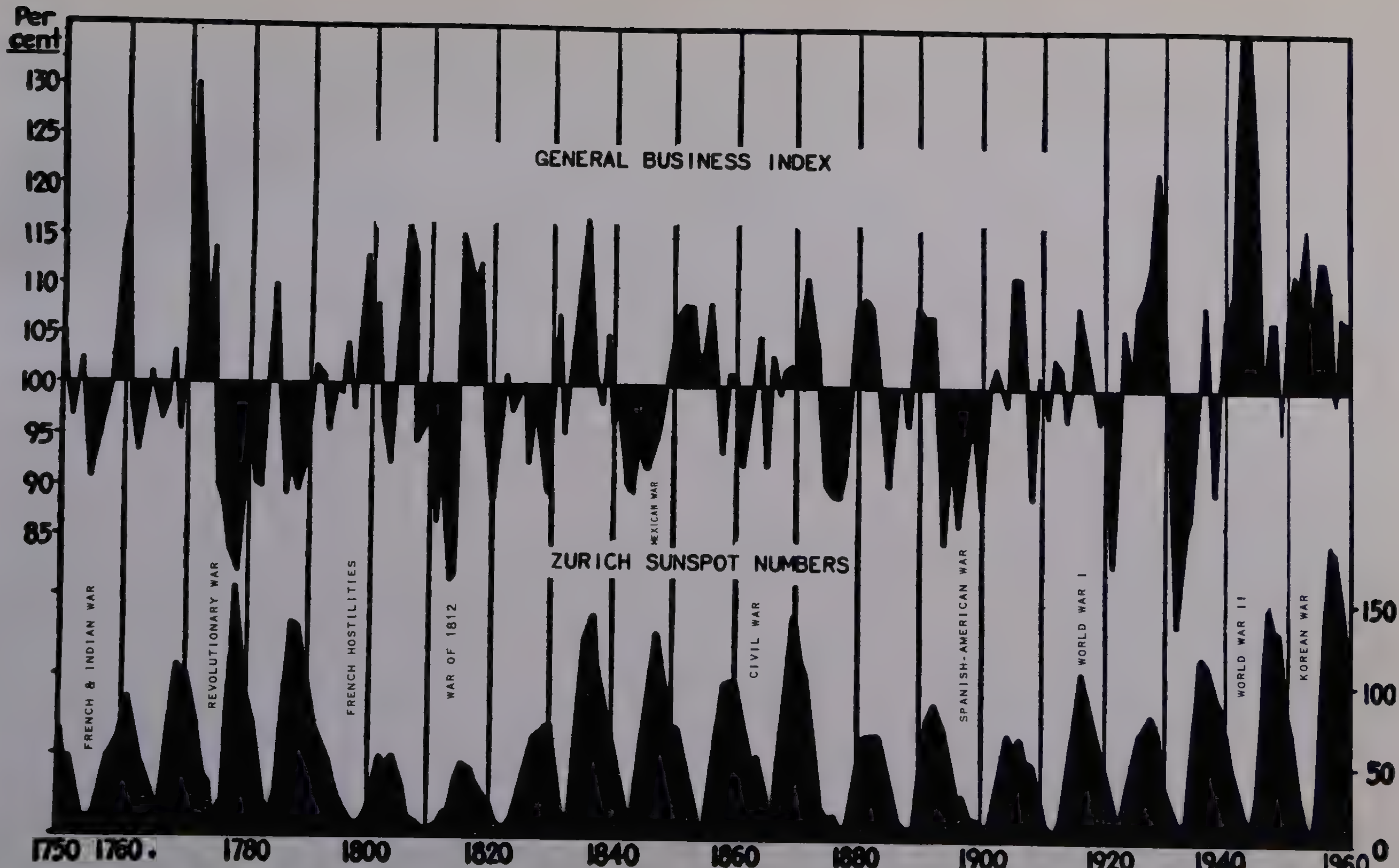


FIGURE 2. Sunspot numbers and business cycles, 1750 to 1960.

Controversial Sunspot Cycle Correlations

Tree rings. Douglas[33] found a persistent correlation between the 11-yr. sunspot cycle and tree rings, using specimens of Arizona pines, the redwoods of California, and the giant Sequoias. He concluded that periods of sunspots have been related to periods of abundant or deficient moisture in the great Southwest. Abetti[34] presents a chart showing 2.8 to 44.0 yr. cycles obtained by F. Vercelli from a periodic analysis of the dendrological sequence of a *Sequoia gigantea* from 274 B.C. to 1914 A.D. Abetti states, "Extremely obvious, however, are the oscillations of about 11.1 years, which must be traced to solar causation, and the way in which the agreement with the curve of relative sunspot numbers is suppressed for a certain number of years, later to be resumed with increasing amplitude." D. J. Schove (in these pages) states, "The solar cycle and the greater fluctuations are reflected in climatic fluctuations noted in evidence (notably from Islamic sources and again received from the Spectrum of Time) relating to tree rings, and to famines and drought in low latitudes."

R. A. Bryson and J. A. Dutton (this monograph), however, were unable to find the 11-yr. sunspot cycle in variance spectra analysis of tree-ring data from the American southwest.

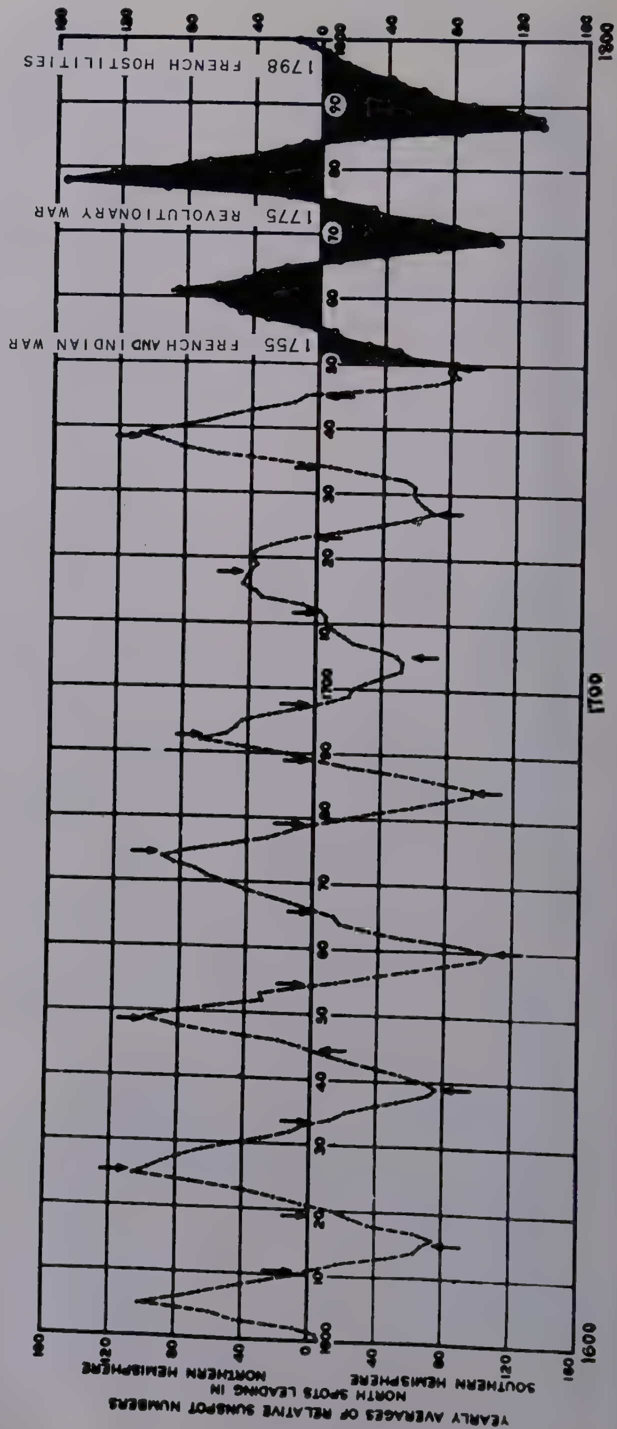
The 11-yr. sunspot cycle. While it is customary to speak of an 11-yr. sunspot cycle, many fail to realize that, as Schove[35] points out, the cycle may vary in length from a minimum of 8 yr. to a maximum of 16 yr. Thus the 11-yr. figure is a mere mathematical abstraction. Although approximations may be obtained by simple averaging, or by some forms of harmonic analysis, Anderson's[32] harmonic analysis of an apparent 312-yr. cycle does not reveal an 11-yr. cycle at all. The strongest component is the 22¼-yr. cycle shown in FIGURE 3.

Various contributors to this monograph obtained different results when analyzing the same or similar data by different mathematical techniques. It should be obvious that a technique that cannot distinguish between so-called cycles in random numbers and real cycles inherent in the data being analyzed is worse than useless. If cycles derived by mathematical techniques are synthetic, they cannot be used for successful forecasting. FIGURE 3 shows that the use of a 22¼-yr. harmonic curve missed the 2 future peaks by 4 yr. in each instance.

Conclusion

Definite correlations have been found between solar activity and geomagnetic activity, auroras, disturbances to telegraph, telephone, radio, overhead and underground electric transmission circuits, weather, and climate. While there is strong evidence for a relationship in tree-ring data and with the Business Cycle, there is some difference of opinion in these fields.

The use of mathematical techniques to derive cycles from data poses the question whether the cycles are not introduced by the particular technique used. Care should, therefore, be exercised in attempting to use such cycles for forecasting purposes.


1798 FRENCH HOSTILITIES
1775 REVOLUTIONARY WAR
1755 FRENCH AND INDIAN WAR
1600
1700
1800
NORTHERN HEMISPHERE
NORTH SPOTS LEADING IN
SOUTHERN HEMISPHERE
YEARLY AVERAGES OF RELATIVE SUNSPOT NUMBERS

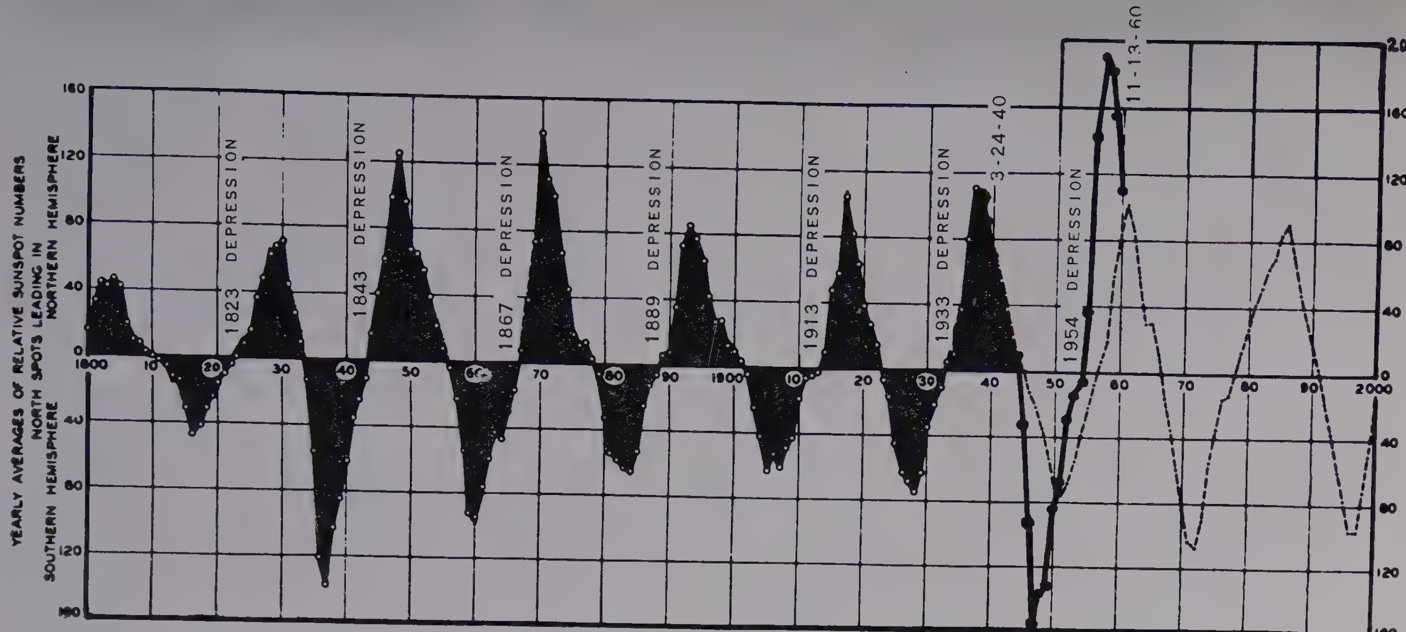


FIGURE 3. Anderson's[32] 22-yr. sunspot cycle. Measured and computed sunspot numbers, 1600 to 2000 A. D. Key: solid line indicates measured values; light dashed line indicates values obtained by adding up the components computed from the measured values; components are harmonics of a 312-yr. period.

References

1. Herschel, W. 1801. Observations of the Sun. Phil. Trans. Roy. Soc. London, England.
2. Schwabe, H. 1844. Periodicitat der Sonnenflecken. Astronomische Nachrichten, XXI, 234.
3. Herschel, J. F. W. 1867. Outlines of Astronomy. : 261. Longmans, Green. London, England.
4. Skilling, W. T. & R. S. Richardson. 1947. Astronomy. : 355. Henry Holt. New York, N. Y.
5. See, T. J. J. 1917. Electrodynamic Wave Theory of Physical Forces. Nichols & Son. Lynn, Mass.
6. Abetti, G. 1957. The Sun. : 296, 301, 305–306. Macmillan. New York, N. Y.
7. Chapman, S. & J. Bartels. 1940. Geomagnetism. **1:** 416. Oxford Univ. Press. London, England.
8. Stetson, H. T. 1947. Sunspots in Action. 74. Ronald Press. New York, N. Y.
9. Anderson, C. N. 1940. Notes on the time relation between solar emission and terrestrial disturbances. : 503. Proc. I.R.E. November.
10. Fraser, R. 1957. Once Round the Sun. : 77. Macmillan. New York, N. Y.
11. Davidson, W. F. 1940. The magnetic storm of March 12, 1940—effects in the power system. E.E.I. Bull. July.
    Davidson, W. F. 1941. Sun-spot disturbances of terrestrial magnetism. Electrical Engineering. February.
12. Germaine, L. W. 1940. The magnetic storm of March 24, 1940—effects in the communication system. E.E.I. Bull. July.
13. Electrical World. 1960. High Sunspot Activity Mars Transmission Operation, Dec. 5.
14. Times. 1960. The New York Times, August 7.
15. Wolf, R. 1852. Trans. Soc. Natl. Phil. Berne, Switzerland.
16. Pogson, N. R. 1858. East India (Report of Famine Commission) Appendix i. Miscellaneous Papers (Blue Book published in 1881). : 32, 33.
17. Lockyer, N. 1878. Evidence concerning the cycle of sunspots and of rainfall in southern India; famine commission of 1878–80, p. 20. Appendix i. Famine Series of Blue Books.
18. Garriott, E. B. 1902. The New York American and Journal, Sunday Edition, August 17.
19. Beveridge, W. H. 1922. Wheat prices and rainfall in Western Europe. J. Roy. Stat. Soc. **85**(3): 412.
20. Clayton, H. H. 1923. World Weather. Macmillan. New York, N. Y.
21. Clough, H. W. 1933. The 11-year sunspot period, secular periods of solar activity, and synchronous variations in terrestrial phenomena. Monthly Weather Rev. April.
22. Clayton, H. H. 1943. Solar Relations to Weather and Life. : 47, 48. The Clayton Weather Service. Canton, Mass.
23. Huntington, E. 1945. Mainsprings of Civilization. Wiley. New York, N. Y.
24. Abbot, C. G. 1946. The sun makes the weather. Sci. Monthly, April.
25. Gillette, H. P. 1946. Weather cycles and their causes. Water & Sewage Works, June.
26. Jevons, W. S. 1875. The solar period and the price of corn. Bristol Meeting. British Assoc. 1875.

27*a*. Jevons, W. S. 1878. The Periodicity of Commercial Crises and Its Physical Explanation. Dublin Meeting, British Assoc. Aug. 19.

27*b*. Jevons, W. S. 1878. Commercial crises and sunspots. Nature. Nov. 14.

28. Jevons, H. S. 1909. The sun's heat and trade activity. Contemporary Review, August.
29. Kitchin, J. 1923. Review of Economic Statistics, January.
30. Garcia-Mata, C. & F. I. Shaffner. 1934. Solar and economic relationships. Quart. J. Economics, November.
31. Silberling, N. J. 1943. The Dynamics of Business. McGraw-Hill. New York, N. Y.
32. Anderson, C. N. 1939. A Representation of the Sunspot Cycle. Bell System Technical Journal, **XVIII:** 292–299, April.
33. Douglass, A. E. 1931. Tree rings and their relation to solar variations and chronology. Ann. Rept. Smithsonian Inst., 304.
34. Abetti, G. 1957. The Sun. : 316–318. Macmillan. New York, N. Y.
35. Schove, D. J. 1955. The sunspot cycle 649 BC to AD 2000. J. Geol. Research. June.

THE PATTERN OF SOLAR CLIMATIC RELATIONSHIPS*

H. C. Willett
Massachusetts Institute of Technology, Cambridge, Mass.

Introduction

Evidence as to the tentative pattern of atmospheric reaction to sudden solar disturbance of the corpuscular (charged particle) and of the ultraviolet types is briefly reviewed in this paper.

Similarity of the reactions of the mean patterns of the general circulation at alternate sunspot maxima to these two sudden reaction patterns is noted, and the principal corresponding climatic effects are briefly noted.

A similar comparison is made of the climatic responses to the 80- to 90-year sunspot cycle, notably as regards the northern hemisphere pattern of the warming trend from 1900 to 1940, which is analyzed in some detail for the winter and for the summer season separately.

Short-Term Reactions of the General Circulation to Sudden Solar Disturbances

The purpose of this paper is to point out certain similarities that exist between indicated short-term reactions of the general circulation of the northern hemisphere to sudden solar disturbance, and some of the apparent long-term climatic reactions to the double (Hale) and to the 80- to 90(89?)-year sunspot cycles. To make this comparison we must start with a brief reference to the classic paper of the Duells,[1] which deals with apparent reactions of the general circulation in the eastern Atlantic and western Europe to sudden outbursts of solar corpuscular and solar ultraviolet radiations. Subsequent analysis by Craig[2] indicates that in just this sector of the northern hemisphere atmospheric reaction is stronger than in any other but, unfortunately, no independent extensive investigation of these relationships has been completed.

Duell and Duell used the international geomagnetic character figures (C_i) as an index of solar corpuscular invasions. They found that the effect of these invasions is most clearly reflected in the day-to-day changes of the sea-level pressure pattern during the winter season and during years of less sunspot activity (annual mean relative sunspot number <40). The restriction of the occurrence of clear patterns of day-to-day pressure change to less active sunspot years may be due to the mutual interference of more frequent particle invasions, or of conflicting ultraviolet outbursts, during the more active sunspot years.

FIGURE 1, selected from the Duells' paper, represents in mb. the average departure from normal of the sea-level pressure over Europe and the northeastern Atlantic Ocean on successive days, as indicated, following a marked geomagnetic disturbance. The normal pressure from which these departures are taken is defined as the average pressure for the months November to February inclusive, for those 16 years of the period 1906 to 1937 for which the relative sunspot number was less than 40. Five days of each of these 64 months

* This study was carried out almost exclusively by the United States Weather Bureau-Massachusetts Institute of Technology Cooperative Extended Forecasting Project.

were selected by the Royal Meteorological Institute of the Netherlands (in De Bilt) on the basis of the international geomagnetic character figures as being most disturbed. These 320 disturbed days constitute the zero days for which sea-level pressure departures were averaged and from which the succeeding

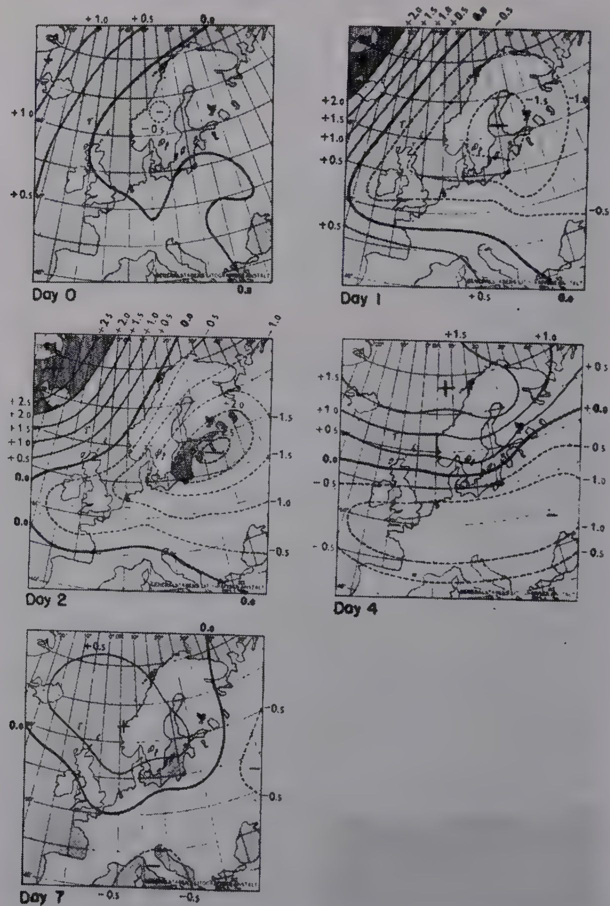


FIGURE 1. Sea-level pressure field from long-period means (in mb.) as related to all those days (320) when the ionosphere was particularly disturbed. Reproduced by permission of *Geografiska Annaler*.[11]

average pressure departures were computed on successive days as indicated in FIGURE 1.

The following points may be remarked in connection with these departure patterns:

The second day after the zero day shows the peak disturbance of the pressure field, in the sense of the southeastward displacement of the prevailing storm

track of the Atlantic toward Europe. In view of the large number of cases included in this sample and of the very modest criterion for a key day, there can be little question as to the statistical significance of the departure patterns of days 1 to 4, in contrast to the flatness of the patterns of days 0 and 7.

The peak effect of the solar corpuscular radiation on the latitudinal (zonal) distribution of pressure apparently is reached on days 3 and 4 after the zero day. In checking the Duell's results on a more extended northern hemisphere grid, Craig[2] verified the essential correctness of the Duell's analysis of the eastern Atlantic and European sector. He established the further fact that on a hemispheric basis the peak effect on the zonal distribution of pressure, reached about 4 days after the zero day, is one of a mean pressure rise of +1.6 mb. at 70° N, and a fall of −0.5 mb. at 40° N, which represents no insignificant weakening of the westerlies in higher middle latitudes.

The claim that this pattern of change does not appear with equal clarity in two samples when the data are broken down by odd and even years does not detract from the significance of these patterns. Many features of the general circulation appear to fluctuate much of the time in a two-year cycle. The pattern must be correspondingly more highly significant in alternate years.

To attempt to evaluate also the effect of sudden solar ultraviolet invasions on atmospheric pressure, the Duells used the occurrence of very bright chromospheric eruptions (solar flares) as an index of ultraviolet invasion. The procedure followed by the Duells is similar to that followed for the solar corpuscular invasions, except that the necessary data were available only for a six-year period (1936 to 1941, inclusive), and 500-mb. contours were selected in place of sea-level pressure as probably more directly responsive to insolational changes. All 51 cases during this period of a very strong flare not preceded within five days by an equally strong flare were used irrespective of season or year.

FIGURE 2, from the Duell's paper, represents the height changes in dkm. of the 500-mb. contours from day −1 to day 0 and to day +1, for all cases and by seasons as indicated. To be remarked in particular in connection with these change patterns are the following facts:

The pronounced insignificance of the change patterns to day 0, in contrast to the much stronger patterns to day +1. A change of slope of >70 dkm. (>230 feet) between the Baltic and western Scotland for 51 cases selected in this manner cannot be dismissed as accidental.

The pattern of change is pronouncedly towards the pressure jump pattern noted by Rossby and discussed by Rex.[3] This occurs typically as blocking anticyclogenesis over Europe in response to strengthening of the zonal westerlies across the Atlantic, with a resultant splitting over Europe of the strong Atlantic jet into a northerly and a southerly branch skirting on either side the blocking center over northern Europe.

The pattern changes for the summer and winter seasons are mutually consistent, with the indicated relative poleward displacement during the summer season of the strengthening of the jet over the Atlantic, as is to be expected.

Craig[2] computed sea-level pressure changes for the northern hemisphere for part of the 51 cases of ultraviolet invasion as selected by the Duells. He generalized his results by saying that the pattern of change tends to be one of

intensifying the initial pattern, that is, for pressure to rise where it is initially high and to fall where it is initially low. This result is entirely consistent with an intensification of the zonal circulation across the oceans.

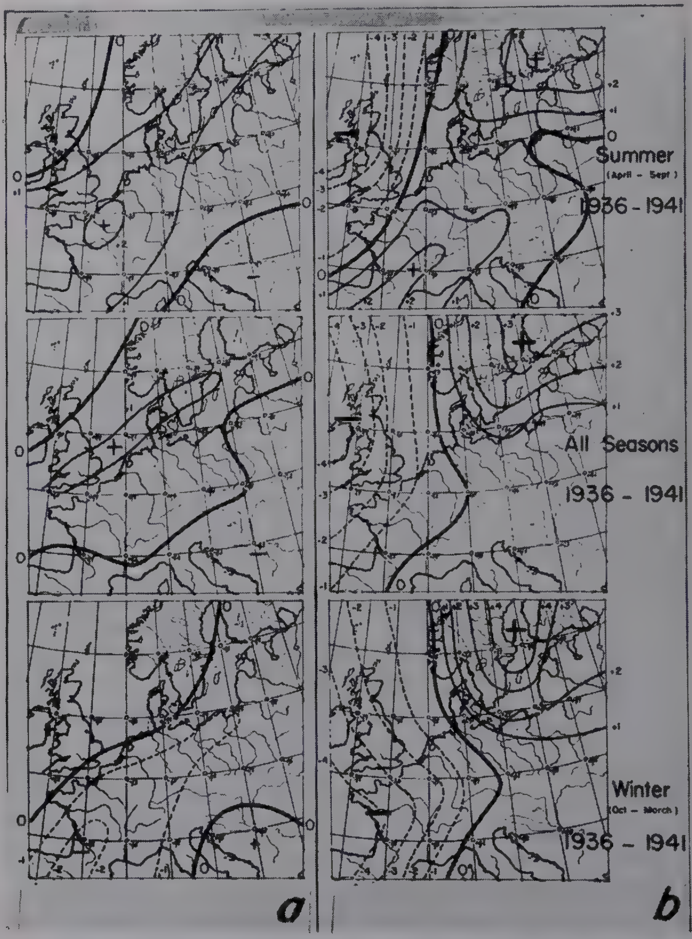


FIGURE 2. The average behavior of the absolute topography of the 500-mb. surface as related to all 51 very intense ultraviolet radiation invasions (between 900 and 1500 GMT) that were not preceded on the previous 5 days by equally strong invasions. (*a*) Average change in dekameters from the preceding day to the invasion day. (*b*) Average change in dekameters from the preceding day to the day that follows the invasion. Reproduced by permission of *Smithsonian Miscellaneous Collections*.[1]

In a detailed analysis of upper winds over the tropical Pacific during a brief period in 1951, Palmer[4] showed that each occurrence of strong flare during this period was followed in 2 or 3 days by a significant increase of the 300-mb. tropical easterlies, hence by inference (and statistics) also of the mid-tropospheric zonal westerlies on the poleward side of the subtropical high-pressure belt. This result represents an additional confirmation, from a very limited

sample of data, of the tendency for strong solar flare (ultraviolet) activity to strengthen zonal circulation across the oceans in middle and lower latitudes.

The Double (Hale) Sunspot Cycle

Although the reality of a double sunspot cycle very legitimately may be questioned on the basis of relative sunspot numbers alone, its physical reality in terms of sunspot magnetic fields, of geomagnetic and of other indices of solar activity cannot be questioned (Willett[5]). Likewise fluctuations of terrestrial atmospheric circulation are more significantly correlated with the 22-year (±2 years) than with the 11-year sunspot cycle.

Hanzlik[6] first pointed out the great difference between the effects of alternate

FIGURE 3. Change of mean winter pressure from sunspot minimum to the following majo sunspot maximum (hundredths of an inch). Reproduced by permission of *Geografiska Annaler*.[11]

sunspot maxima on the distribution of sea-level pressure over Europe, particularly during the winter season. Willett[7] extended Hanzlik's analysis through two additional sunspot maxima (through 1937 to 1939), and over all of the globe for which data were available.

FIGURES 3 and 4 are slightly smoothed reproductions of 2 figures from Willett's paper. They represent, respectively, in hundredths of an inch, the change in mean winter sea-level pressure from the 3 years centered on a sunspot minimum to the 3 years centered on the following *major* sunspot maximum and, similarly, from a sunspot minimum to the following *minor* sunspot maximum. The analysis is based on data from *World Weather Records*[8] for all stations having a record of sea-level pressure extending through more than 2 of the 4½ double sunspot cycles included between the minimum of 1842 to 1844 and the maximum of 1937 to 1939. For stations with a complete record FIGURE 3 represents change patterns between means for 15 winter seasons, FIGURE 4

between means for 12 winter seasons and, in no case, between means for less than 6 winter seasons. The record was complete for the entire period only at a few stations over northern Europe, but in general the record was longest in just that region of Europe and the north Atlantic where the change patterns are strongest.

In connection with FIGURES 3 and 4 the following comments are worth citing:

The impressive reversal of the change of the winter pressure gradient between the British Isles and southeast Greenland going into alternate sunspot maxima, amounting to about a 10 mb.-change of the gradient at the maximum point. Since these alternate periods of change in the double sunspot cycle are separated

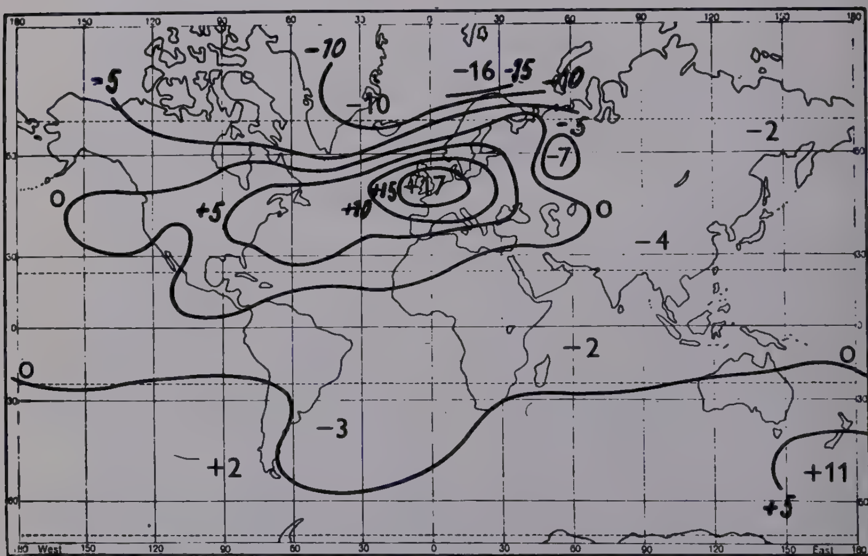


FIGURE 4. Changes of mean winter pressure from sunspot minimum to the following minor sunspot maximum (hundredths of an inch). Reproduced by permission of *Geografiska Annaler*.[11]

by equally long intervening periods, they are relatively independent; hence one cannot be considered merely as a relaxation of the pressure pattern from the other.

The striking similarity of FIGURE 3 to the peak change of pressure following a solar corpuscular invasion (cf. day 2, FIGURE 1). FIGURE 3 indicates an average increase of pressure by about 2 mb. at 70° N, and a decrease of about 0.5 mb. at 40° N, almost identical with the values determined by Craig for the third or fourth day following a solar corpuscular invasion. The implication of this similarity is that solar corpuscular activity dominates the influence of the major sunspot maximum on the circulation pattern. In this connection it may be remarked that the highest level of geomagnetic disturbance, presumably reflecting solar corpuscular activity, does fall exactly in the 3-year period of the major sunspot maximum.

Similarity of FIGURE 4 to the change of circulation following an ultraviolet invasion (FIGURE 2). Although a strict comparison of a pressure change pattern at sea level with one at 500 mb. is not meaningful, it is quite clear from FIGURE 4 that this change pattern does represent a marked intensification of the zonal westerlies at higher latitudes across the north Atlantic. Furthermore, with moderate ridging indicated at the surface east of the Baltic, as well as an intensified flow of warm Atlantic air into this region, this pattern of change indicates even stronger ridging east of the Baltic at 500 mb. Since these two features stand out from FIGURE 2 as principally characteristic of the reaction of the general circulation to solar ultraviolet invasions, it is suggested that the effect of the minor sunspot maximum on the general circulation is predominantly one of enhanced solar ultraviolet rather than of corpuscular activity. In this connection it may be remarked that geomagnetic disturbance is relatively low at the minor sunspot maximum, reaching its peak during the 3-year period following the 3 years of the minor sunspot maximum.

An important climatic consequence of the relatively zonal character of the general circulation during the minor, in contrast to the major maximum phase of the double sunspot cycle, is indicated also by the summer drought pattern. The zonal pattern should favor relatively maritime conditions, as opposed to relatively continental, in the interior of continents in middle and lower middle latitudes. All of the prolonged drought periods in our western plains have fallen in the major maximum decades, that is, in the nineties, the teens, the thirties, and the fifties. The intervening decades have been free of prolonged drought. There is some evidence that a similar drought pattern has been followed also in the southern Union of Soviet Socialist Republics and in Australia.

The 89-Year(?) Sunspot Cycle

Significant worldwide changes of climate appear to have been associated with this cycle of solar activity (Willett[9]). It is not proposed in this discussion to explore the complete pattern of these solar-climatic relationships at length, but rather to look closely at just one phase of the pattern.

The 80- to 90-year cycle of solar activity is defined primarily by a sharp break from peak sunspot number during the last quarter of the cycle to minimum solar activity during the first quarter of the following cycle. During the second and third quarters sunspot activity increases gradually and erratically, with some suggestion of a 40- to 45-year cycle, reaching final peak activity again late in the fourth quarter.

The corresponding pattern of climatic change manifests its sharpest break, from a low index blocking cellular stress pattern during the fourth quarter of peak sunspot activity, to a low-latitude zonal pattern of essentially cool maritime climate in middle and lower latitudes during the first quarter of minimum sunspot activity. During the second and third quarters of irregularly increasing sunspot activity, the zonal climatic pattern tends to shift poleward, with a warming trend in most latitudes. The weakening of the westerlies in lower middle latitudes that results from their poleward shift and strengthening in higher latitudes causes an intensification of the normal climatic continental-maritime seasonal contrasts equatorward of 50° N. This phase of the climatic

change is irregularly variable from one cycle to the next, corresponding to the change of sunspot activity. The high-latitude zonal pattern of mild climate breaks down into the blocking low-index climatic stress pattern during the fourth quarter of peak sunspot activity.

During the last three cycles the fourth quarter of maximum sunspot activity and of climatic stress, preceding the sharp break to the first quarter of minimum sunspot activity and of cool moist conditions in middle latitudes, reached its peak with the minor sunspot maximum of 1787, with the major maximum of 1871 and, presumably, with the major maximum of 1957. The current cycle appears to have resembled much more closely in detail that cresting in 1787 than that cresting in 1871, both solarwise and climaticwise.

The outstanding climatic change of recent decades has been the much-discussed warming trend of the past 40 or 50 years, noted most strikingly in the winter temperatures of the higher latitudes of the northern hemisphere. Since the documentation of other changes possibly equally impressive during the two previous 80- to 90-year cycles is completely inadequate for similar hemispheric analysis, the further discussion is directed primarily at the pattern of this particular climatic trend, which reached its peak from the second to the third quarter of the cycle now drawing to a close, in a pattern that appears to be strictly analogous to the sequence of events at the same phase of the second cycle preceding.

The first attempt to establish the synoptic pattern of this warming trend was made in a study by Willett[10] with the assistance of J. Murray Mitchell. This study showed that by latitude zones of the northern hemisphere, much the strongest warming occurred quite sharply poleward of 60° N from the period 1900-1919 to 1920-1940, that is, from the second to the third quarter of the 80-year cycle. Equatorward of 60° N the trend was smaller and more erratic, but in most latitudes the change was most significant between these 2 periods, particularly so between 10°–30° N where, in terms of standard deviations, it probably was quite as significant as in the 60° to 80° zone.

The above study by Willett[10] and Mitchell attempted to establish the geographical pattern of the winter warming trend between these two periods, by making use of selected stations from *World Weather Data*.[8] FIGURE 5, taken from Willett's paper, shows the geographical pattern of that warming in *tenths* of °F. To be noted in particular in this figure are:

The large degree of warming, centered near Spitzbergen, Norway, and extending across the north Atlantic from Greenland to extreme northern Europe and eastward into Siberia.

The warming in Alaska and northwestern Canada, apparently recurring in a downstream crest of the zonal westerlies in the eastern United States.

Extensive cooling across central Eurasia, and less extensive cooling on either side of the warm crest in the eastern United States.

The effort that was made in this analysis to link the continental areas of cooling with an arctic source of advection was based on hypothesis in the absence of data. More complete data discussed below suggest that this linkage may be at least partly erroneous, that the areas of continental cooling perhaps are largely cut off from an arctic source of cold air.

The strong pattern of change of temperature over the great Eurasian land

mass, and the weaker pattern of change over the smaller North American land mass, both suggest the influence of an effective increase of the winter-season continental influence from the first to the second 20-year period. Increased anticyclonic circulation, or upper level ridging, on the western side of these land masses, more or less proportional to the area of the land mass, appears to be primarily conducive to the observed pattern of temperature change. The continental character of the pattern of temperature change was supported by the further fact that the comparatively weak pattern of change of the annual mean temperature of the northern hemisphere suggested some opposite continental effect during the warm season, but no analysis of the change for the summer season separately was prepared.

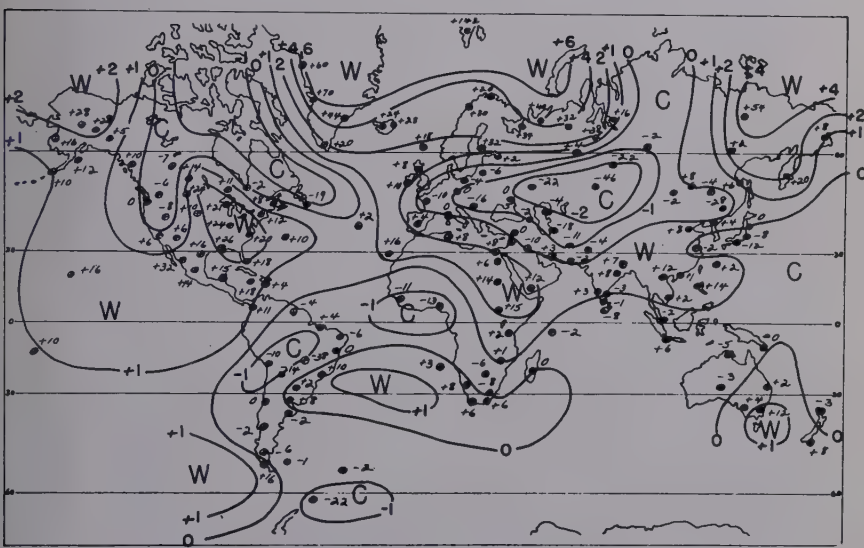


FIGURE 5. Pattern of temperature change, °F. Decades 1920 to 1939 minus decades 1900 to 1919. Reproduced by permission of The Royal Meteorological Society, London, England.[12]

Since data of a sort were available for the purpose, it was decided to take a quite independent approach to the determination of the patterns of change of circulation and temperature over the northern hemisphere from the 1900-1919 to the 1920-1939 period, for the winter and for the summer calendar seasons separately. For this purpose northern hemisphere maps of monthly mean sea-level pressure and of 500-mb. contours were used. The sea-level pressures are derived from the United States Weather Bureau northern hemisphere map series. The 500-mb. means are those computed by Sam Solot, based on the sea-level mean data extrapolated with the aid of all available upper air data and a careful consideration of all available temperature data and probable lapse rates. Whatever degree of confidence one may have in their reliability, taken in conjunction with the sea-level mean maps they probably contain the most complete information available about the monthly mean temperatures of the lower half of the atmosphere of the northern hemisphere

from 1900 to 1939. At least the pattern of change of temperature derived in this manner for the northern hemisphere can be compared with that derived independently from Clayton's data. Unfortunately Solot's data do not extend equatorward of 25° N.

FIGURES 6 to 8 for the winter season, and FIGURES 9 to 11 for the summer season, contain respectively the changes from the 1900–1919 period to the 1920–1939 period, of the seasonal means of the sea-level pressure, of the 500-mb. contour heights and of the mean virtual temperature defined between these

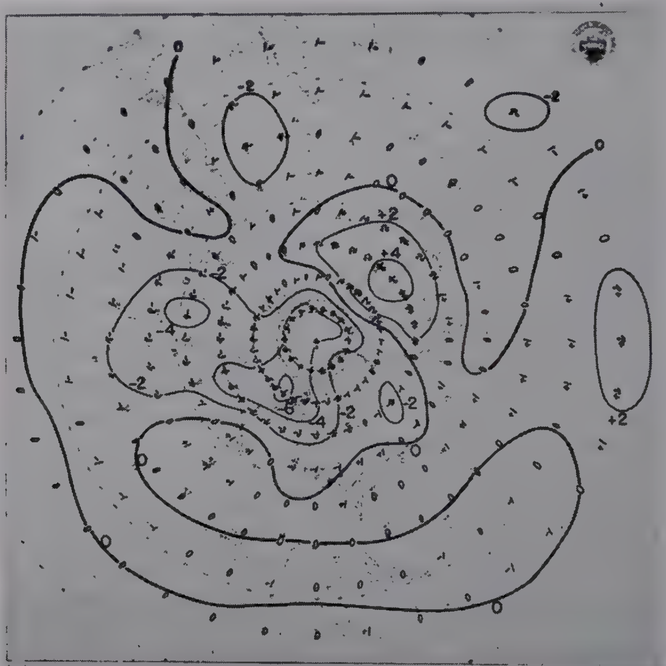


FIGURE 6. Winter season: sea-level pressure, decades 1920 to 1939 minus decades 1900 to 1919.

levels. The remainder of the discussion is concerned with these six change patterns.

FIGURE 6 presents the change of the winter mean sea-level pressure in millebars from the first to the second 20-year period (second to third quarter of the 80- to 90-year sunspot cycle). Note in particular the following points:

Marked intensification of the high latitude zonal westerlies, particularly along the northwestern edges of the two major continental land masses in higher latitudes. This intensification of the zonal westerlies is forced over by the larger Eurasian land mass to the North American side of the Pole. Between 60° to 80° N it represents an increase of the mean sea-level northern hemisphere (N. H.) zonal westerlies of 0.98 m./sec. Note the general similarity across the north Atlantic to FIGURE 4.

Strong anticyclogenesis centered east of the Baltic (northwestern Eurasia) as indicated by the temperature change pattern of Willett's earlier study. At least relatively some slight anticyclogenesis also on the northeast coast of Asia.

Weak anticyclogenesis centered on the west and on the east coasts of North America, at somewhat lower latitudes than in Asia.

FIGURE 7. Winter season: 500 mb. 10′ units (decades 1920 to 1939 minus decades 1900 to 1919).

Pronounced deepening, or cyclogenesis, extending in a trough from the western north Pacific across the pole to the western north Atlantic. Some tendency to secondary troughing in east-central North America and inland over eastern Asia.

The entire northern hemisphere pattern of change of the sea-level pressure may be expressed as an intensification of the circumpolar zonal circulation in higher latitudes centered on the North American side of the pole, with in-

creased maritime-continental pressure contrasts in middle latitudes over North America and in higher middle latitudes over Asia.

FIGURE 7 presents the change of the winter-season mean 500-mb. contour heights in tens of feet from the first to the second 20-year period. In general it

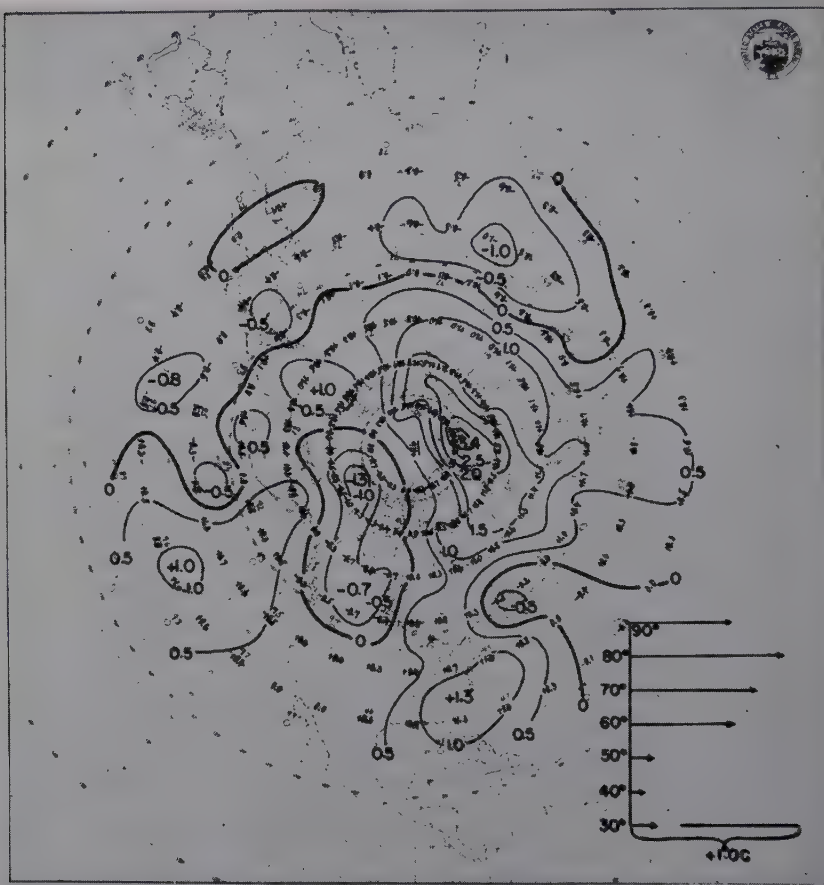


FIGURE 8. Winter season: T_v °C. sea level-500 mb. (decades 1920 to 1939 minus decades 1900 to 1919).

supports every major feature of the pattern of change of the sea-level pressure, notably as follows:

Intensification of the circumpolar vortex in higher latitudes displaced to the north American side of the pole. This change of the mean (N. H. 500-mb. zonal westerlies amounts to +1.5 m./sec. between 70° to 80° N, and −1.0 m./sec. between 40°–60° N. At sea level there is a change of −0.3 m./sec. in the mean N. H. zonal westerlies from 50° to 20° N.

The same change of the continental-maritime trough and ridge pattern is

indicated by the sea-level pressure changes, except that over the northeastern quadrant of both continents height rises extend into areas of surface pressure fall. This indicates in both regions that the increased eastward flow of warm air aloft in response to the intensified zonal circulation tends to displace or eliminate at upper levels areas of weak decrease of sea-level pressure.

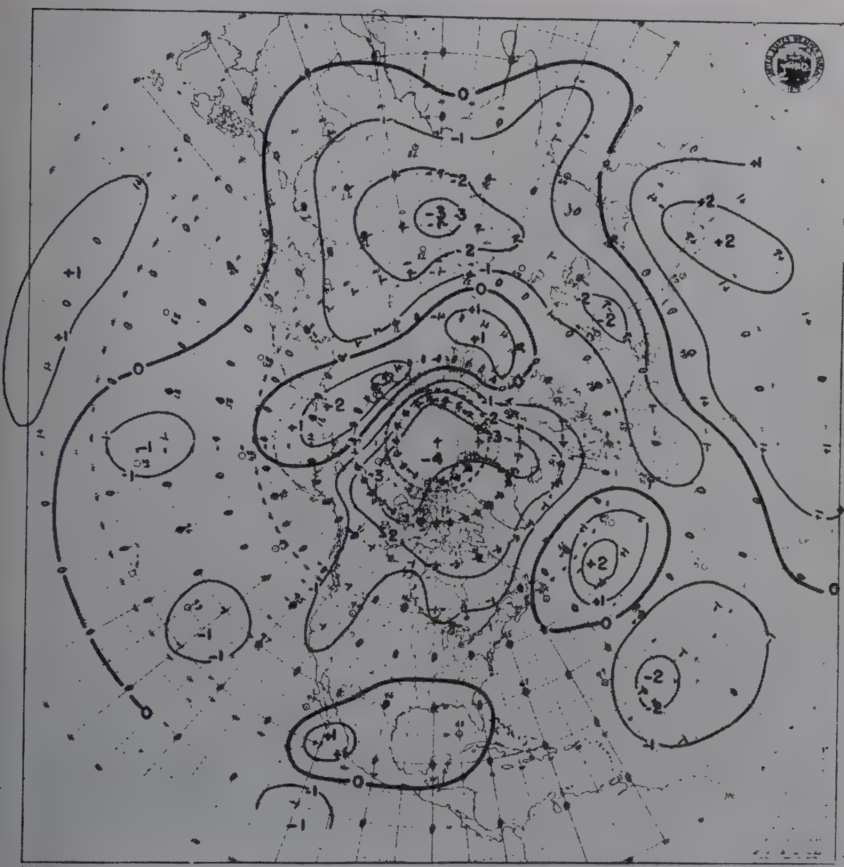


FIGURE 9. Summer season: sea-level pressure (decades 1920 to 1939 minus decades 1900 o 1919).

Particularly noticeable across the continent of Asia is the tendency to weaken he 500-mb. westerlies in middle latitudes at the same time that they are trengthened in high latitudes. This continent contributes most of the change f strength of the N. H. zonal winds noted above.

FIGURE 8 presents the change of the winter season mean virtual temperature etween sea level and 500 mb. in tenths of degrees C. from the first to the sec- nd 20-year period. The profile in the lower right hand corner of the figure

represents the average change of this T_v on each 10° latitude circle from 30° N to the pole. The following facts should be noted:

Every important feature of FIGURE 5, based on *World Weather Records* of station data, is reflected albeit with an expected reduction of amplitude, by these T_v data.

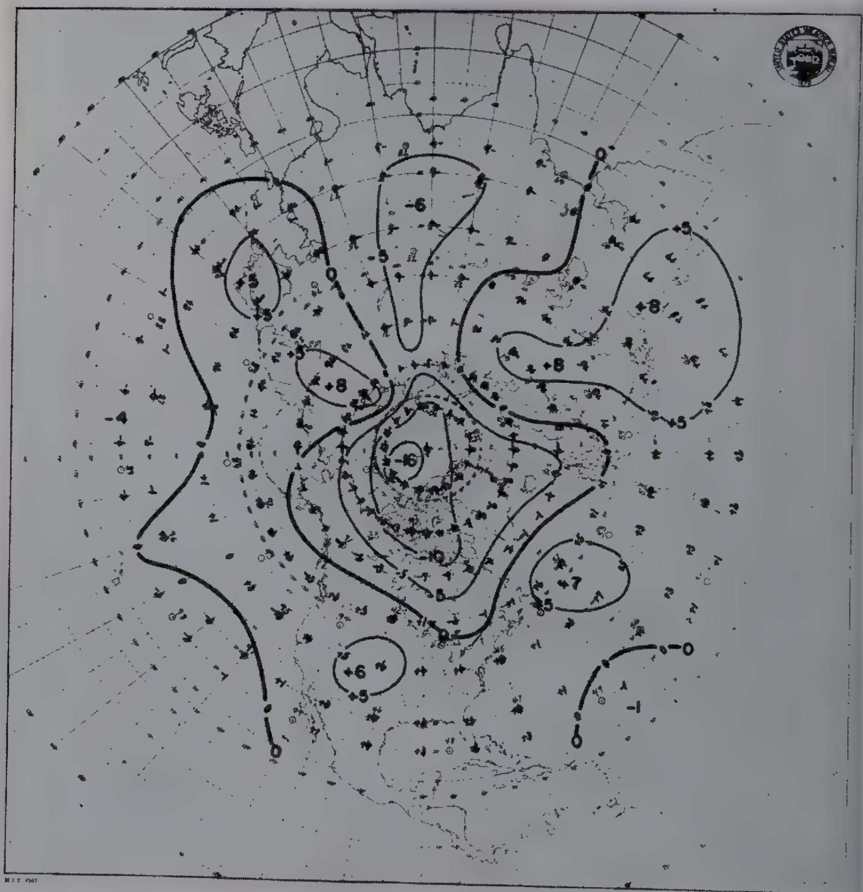


FIGURE 10. Summer season: 500 mb. 10′ (decades 1920 to 1939 minus decades 1900 to 1919).

These data confirm the fact suggested by the 500-mb. contour changes of FIGURE 7, that the surface cooling indicated in FIGURE 5 at Newfoundland, as well as the cooling across all of south-central Eurasia, are both completely disconnected from any advective Arctic source. Quite the contrary condition seems to be true in western North America, however; toward this area there exists the principal tendency for the displacement of the center of the cold circumpolar vortex. In other words, increased continentality during the winter

season tends to favor direct outbreaks of cold Arctic air into western North America, but to discourage them over northeastern North America and generally throughout Eurasia.

The latitudinal pattern of the change of temperature, like that of circulation, obviously is dominated by the Eurasian continent. It is thus reflected pri-

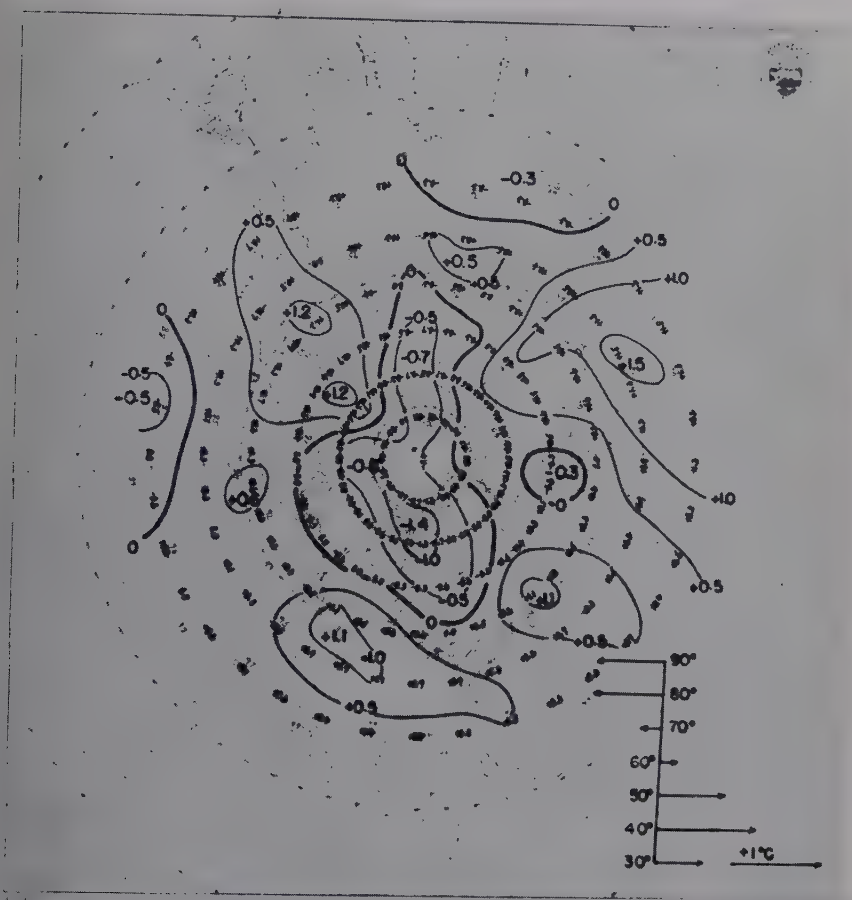


FIGURE 11. Summer season: T_v °C. sea level-500 mb. (decades 1920 to 1939 minus decades 1900 to 1919).

marily by a maximum increase of temperature in very high latitudes, representing essentially greatly increased advection of maritime air along the northern edge of the Eurasian land mass by the strongly intensified zonal westerlies. Minimum increase of temperature is found at 40° N, reflecting the continental cooling in the zone of decreased westerlies across south central Eurasia. It is suggested that the increase of the temperature rise that begins to appear at 30° N and by Clayton's data is indicated to reach a highly signifi-

cant maximum between 10° to 20° N, probably reflects an effective increase of insolational heating with the steadily increasing level of solar activity from the second to the third quarter of the current 80-year cycle. It cannot be decided at this point whether this effective increase of insolational heating represents an increased contribution of thermal energy by solar activity, or merely an effect on the radiational transmissive properties of the atmosphere. The first alternative is preferred by the writer.

FIGURE 9 presents the change of the mean sea-level pressure of the summer season from the 1900-1919 period to the 1920-1938 period. The following features of this figure should be remarked:

A marked intensification of the circumpolar cyclonic vortex centered on the North American side of the Pole strictly analogous to that shown by FIGURE 6 for the winter season, including a similar pronounced weakening of the middle latitude westerlies across the Eurasian continent. It is perhaps remarkable that this major change in the circulation is so uniform for the two opposite seasons. The change of the mean N. H. sea-level zonal westerlies for the summer season is +0.6 m./sec. poleward of 60° N, and −0.5 m./sec. between 50° to 40° N. The negative change in middle latitudes derives entirely from the Eurasian continent.

A tendency towards reversal of the continental-maritime pattern of cyclogenetic and anticyclogenetic change from the winter pattern of FIGURE 6 at every point, over North America, over Eurasia, and over the Atlantic and the Pacific Oceans. In particular the winter pattern of negative pressure changes over the western side of the two oceans is replaced by a summer pattern of pressure rises in the same areas. Consequently the major axis of pressure fall across the pole from the western Pacific to the western Atlantic in winter is rotated to a position from eastern Asia to eastern North America in summer. Likewise pressure falls replace pressure rises over northern Europe.

FIGURE 10 representing the change of the circulation pattern at the 500-mb. level for the summer season shows a cyclonic intensification of the circumpolar vortex, centered on the North American side of the pole even stronger than that for the winter season (FIGURE 7). The increase of the seasonal mean N. H. 500-mb. zonal westerlies averages 1.5 m./sec. from 50° to 80° N, and a negative change of −0.6 m./sec. from 40° to 30° N. At the same time the comparison of FIGURE 10 with FIGURE 7 shows even more clearly at 500 mb. than does the comparison of the sea-level pressure pattern of FIGURE 9 with that of FIGURE 6 the perfect reversal from the winter to the summer season of the monsoonal features of the patterns of circulation change between the 1900-1919 and the 1920-1939 periods. This applies equally to the shift of the relative cyclogenetic troughs from the western to the eastern sides of the two oceans, to the shift of the relative anticyclogenetic ridges from the west coasts of the North American and Eurasian continents well inland, and to the replacement of the ocean-to-ocean axis of relative cyclogenesis across the pole by the continent-to-continent axis as the primary one. In other words, the change of the summer circulation patterns between these periods shows identically the same intensification of zonal circulation in high latitudes and of monsoonal circulation in middle latitudes as does the change of the winter circulation patterns, allowing for the normal seasonal reversal of the continental-maritime influence.

FIGURE 11 represents the pattern of change of the mean T_v between sea level and 500 mb. for the summer season from the first to the second period, to be compared with FIGURE 8 for the winter season. The striking feature brought out by a comparison of these 2 figures is the expected seasonal reversal at every point of the intensified maritime-continental influence on the patterns of change. Note particularly the rise of temperature that occurs over all the major land masses in lower middle latitudes. Comparing the latitudinal profiles of change of temperature in the lower right corner of FIGURES 8 and 11, we see that whereas in winter the minimum average increase of temperature occurs at 40° N, in summer the maximum increase occurs there. In fact, on this parallel not a single meridian shows a negative change for the summer season. This fact is taken as further evidence of the direct insolational contribution at low latitudes to the increase of temperature from the second to the third quarter of the 80- to 90-year solar cycle, the contribution that appears strongest in winter between 10° to 20° N.

Striking also is the complete disappearance of the great winter increase of temperature from Spitzbergen eastward along the northern border of Eurasia. The advection of maritime conditions across northern Siberia that produces the large temperature rise in winter continues also in summer, but naturally the thermal effect in summer is quite different. Note on the temperature change profiles the complete reversal at all latitudes of the profile pattern between the winter and the summer seasons.

By way of a summary conclusion it may be stated that the seasonal patterns of change of the northern hemisphere circulation and of the distribution of temperature from the second to the third quarter of the currently terminating 80- to 90-year climatic cycle verify completely the expectations based on the statistical analysis of past solar-climatic relationships. Both patterns reflect clearly the effects of increased zonal westerlies in the high latitudes, and the intensification of the normal seasonal monsoonal contrasts in middle and lower latitudes. Furthermore, this pattern of change is closely analogous to that following sudden solar ultraviolet outbursts, and to that passing from a sunspot minimum to a minor sunspot maximum. This fact suggests that the dominant characteristic of the trend of solar activity from the second to the third quarter of the 80+-year sunspot cycle may be towards increasing ultraviolet emission (flare activity?).

References

1. DUELL, B. & G. DUELL. 1948. The behavior of barometric pressure during and after solar particle invasions and solar ultraviolet invasions. Smiths. Misc. Coll. **110**(8).
2. CRAIG, R. A. & D. HAWKINS. 1951. Atmospheric pressure changes and solar activity. Special Rept. No. 38, A. M. C. Contract W19-122ac-17.
3. REX, D. F. 1950. Blocking action in the middle troposphere and its effects upon regional climate. Tellus. **2:** 196–211, 275–301.
4. PALMER, C. E. 1953. The impulsive generation of certain changes in tropospheric circulation. J. Meteorol. **10**(1).
5. WILLETT, H. C. 1960. Long-Term Indices of Solar Activity. Scientific Report No. 1, NSF Grant 5931, Cambridge, Mass.
6. HANZLIK, S. 1931. Der Luftdruckeffekt der Sonnenfleckenperiode für die Monate Dezember—, II Mitt., Gerlands. Beitr. z. Geophys., 29.
7. WILLETT, H. C. 1949. Report of the Weather Bureau—M. I. T. Extended Forecasting Project. Cambridge, Mass.

8. Clayton, H. H. & F. L. Clayton. 1944 & 1947. World Weather Records. Smithsonian Misc. Collections. 79, 90, 105, Smithsonian Inst. Washington, D.C.
9. Willett, H. C. 1951. Extrapolation of sunspot-climate relationships. J. Meteorol. (1).
10. Willett, H. C. 1950. Temperature trends of the past century. Centenary Proc. Roy. Meteorol. Soc. : 195–206.
11. Willett, H. C. 1949. Solar variability as a factor in the fluctuations of climate during geological time. Geografiska Annaler. **31**: 295–315. C. C. Wallén, publishers. Stockholm, Sweden.
12. Willett, H. C. 1950. Temperature trends of the past century. 1: 199. Centenary Proceedings of the Royal Meteorological Society. London, England.

SOLAR CYCLES AND THE SPECTRUM OF TIME SINCE 200 B.C.

Derek J. Schove
St. David's College, Beckenham, Kent, England

Introduction

The so-called Spectrum of Time project arose from appeals at international conferences of various kinds (meteorological, geographical, archaeological, and historical) for information about datable climatic and meteorological events. The items received were cross-dated and diagnosed, and those relating to sunspots and the auroras were singled out for a separate study of the sunspot cycles in the historical period. The method was described in various journals (Schove, 1947, 1951*e*, 1950*d*, and 1951*c*), and the detailed results have been tabulated in the *Journal of Geophysical Research* (Schove, 1955*b*). The meteorological information was likewise separated in order to survey the climatic fluctuations of the past 2000 years. Although it was essential to keep these two investigations apart—in order to avoid even unconscious bias—some of the main results will be summarized here. In my two papers included in this monograph, the detailed evidence that has been or is being described elsewhere will be omitted, and the emphasis will be on oscillations and on decadal data of potential significance in the American continent.

Solar Cycles Since 200 B.C.

The dates of the sunspot maxima in the 11-year cycle were normally easy to determine and thus have been summarized in a brief table (Schove, 1956*a*). Observations of the aurora were especially detailed in China, Japan, and Korea, and there were many accounts of portents and visions in Europe that were in fact of auroral origin. Mexican and Red Indian chronicles were also used for Spectrum of Time purposes although no certain auroral identifications were made. It was noticed (Schove, 1948*a*) that before 800 A.D. the last two digits of the years of maximum often followed the "11" timetable (that is, 311 A.D. or 511 A.D.) but that after 800 A.D. the minima often obeyed a similar rule (that is, 1944). The remainder or the "phase" of a particular maximum was defined as the difference between the actual year and the previous year with its last two digits exactly divisible by 11.

The phase—in this sense of remainder—of the various estimated sunspot minima is illustrated in the lower curves of FIGURE 1, and it is hoped that those remainders that could not be determined precisely by the documentary evidence may ultimately be determined from the radiocarbon content of dated tree rings, using the principles described by Stuiver (in this monograph).

The latest maximum in 1957 has a remainder of only 2, and past instances suggest that it is ahead of schedule because of the recent extreme activity. Indeed the next minima might well be delayed until the zero-remainder date of 1966. If this happens, we must expect much less activity in the subsequent cycle, a cycle associated with a late maximum about 1972, in accordance with the empirical formula:

$$\text{Max. year–previous min. year} = 7 - (0.03)n$$

where n is the sunspot number at maximum (Schove, 1955*b*).

The intensity of the historical maxima could be estimated from the documentary evidence. Thus in active solar cycles auroras were both more extensive and more frequent, although the big displays took place occasionally as late as 6 years after the maximum. It was necessary to consider also the variability of the record. For instance, all spectrum-of-time information was very weak in the 3rd century; on the other hand, Chinese nocturnal observations of different phenomena were particularly detailed in the 11th century

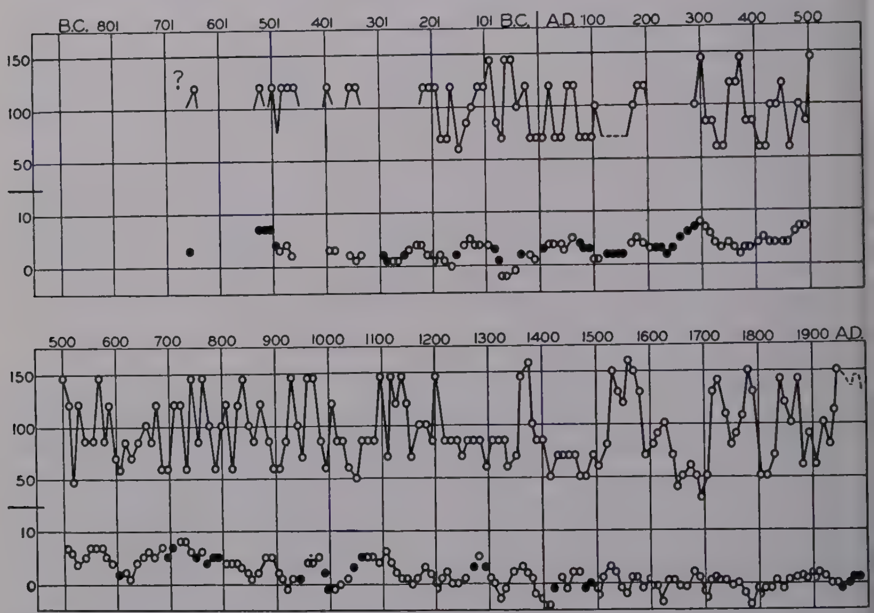


FIGURE 1. The sunspot cycle, 649 B.C. to 2000 A.D. Curves showing intensity and phase, 650 B.C. to 2000 A.D. Upper curve: auroral intensity (sunspot numbers from 1750 A.D.) of maxima. Lower curve: phase (remainder) of minima; ○ = reliable, ● = uncertain. Reproduced by permission of the *Journal of Geophysical Research* (Schove, 1955).

(Schove and Ho, 1958). The intensity of each maximum was expressed as an auroral number—intended to be comparable with the sunspot number—and these are the figures indicated in the upper curves.

The mean intensity was also estimated for each decade, and these decadal values are now tabulated for ease of comparison with meteorological data. It was not anticipated in 1955 that the major fluctuations in auroral activity would correspond to fluctuations in the radiocarbon content of tree rings (noted in this monograph both by Stuiver, and by Willis, and first pointed out to me by R. W. Fairbridge), but the probable errors given in the full table (Schove, 1961) will make it possible in the future to determine whether this similarity is accidental or significant.

Other results of geophysical interest can be determined from the documentary

information. Naked-eye sunspots and other aurorally productive regions traveled around the sun in the 12th century in 27⅔ days just as they do today. The direction of magnetic north was different, but the changes in the meridian can be gauged from the recorded changes in direction of auroral displays. Thus in England in the early 12th century holy men walked due north toward the celestial light, whereas in China the contemporary observed change of auroral direction from NW to N (Schove and Ho, 1958) is not inconsistent with the new paleomagnetic evidence (Watanabe, 1959) from pottery in nearby Japan.

The decadal data may prove useful in testing the significance of the cycles briefly noted in 1955 (p. 141–2). Vague, interrelated cycles were noticed in both period and activity. The 80-year cycle (80 to 90, according to Willett, and 89, according to Visser, 1959; see Anderson's 86-year cycle in these pages) in sunspot length noted by Gleissberg (1958) seemed so real since 1610 that it was taken into account in estimating the dates of some earlier minima, but my adoption of Gleissberg's value of 78 years is perhaps arbitrary. It is possible that if maxima with probable errors of 3 or more are excluded from the analysis, as originally suggested (Schove, 1955*b*, p. 137), the true length and significance of some cycle between 65 and 90 years can be determined. The cycle of about 165 years noted since 1510 might seem double the Gleissberg cycle, but no cycle of this length was noticed in the medieval period. Moreover the Gleissberg cycle did not seem to be clearly marked in the intensity pattern where, on the other hand, a 200-year cycle seemed significant (Dewey, 1958). Solar activity was apparently greater in even centuries (such as the present 20th century), although this rule was clear only since 300 A.D.

Climatic Fluctuations in Europe since 1100 A.D.

The Spectrum of Time information has already proved useful for the study of historical meteorology both in Europe and in Asia. In Europe it is now possible to estimate changes of pressure and pressure gradient through the centuries (Schove, 1961, 1953*e* and *f*). From 1800 to 1950 the barometric evidence is sufficient (Schove, 1958, 1961) to determine the pressure (top curve of FIGURE 2), at a fixed central position ("The Wash") in eastern England, and to determine also the changes in the strength (central curve) and direction (lower curve) of the prevailing wind.

The pressure chart for the modern period shows maxima about 1830, 1860, 1900, and 1945, as if the barometer were in step with the mathematical series 3, 6, 10, and 15. These pressure peaks were studied geographically (FIGURE 3), and were found to have moved from north to south according to the principle of south steering (Schove, 1950*a* and *b*). Temperature patterns by 30-year periods moved southward in the same way (FIGURE 4), the cold phase (T–min.) preceding the pressure peak (P–max.) by about a decade (see FIGURE 5).

Terms such as P–min. and T–min. will be used again in the course of this paper; the following typical definitions are therefore listed here for convenience: S–max., 30-year period of maximum sunspot/auroral activity (dated by the 15th year); t–max. (cf. p–max., r max.), 10-year period of maximum temperature (dated by the 5th year); and f(R)–max. (cf. f(T)–max.), 30-year period of maximum tree growth, where rainfall is the presumed major factor.

Use of this climatological algebra enables us to study the migration of pres-

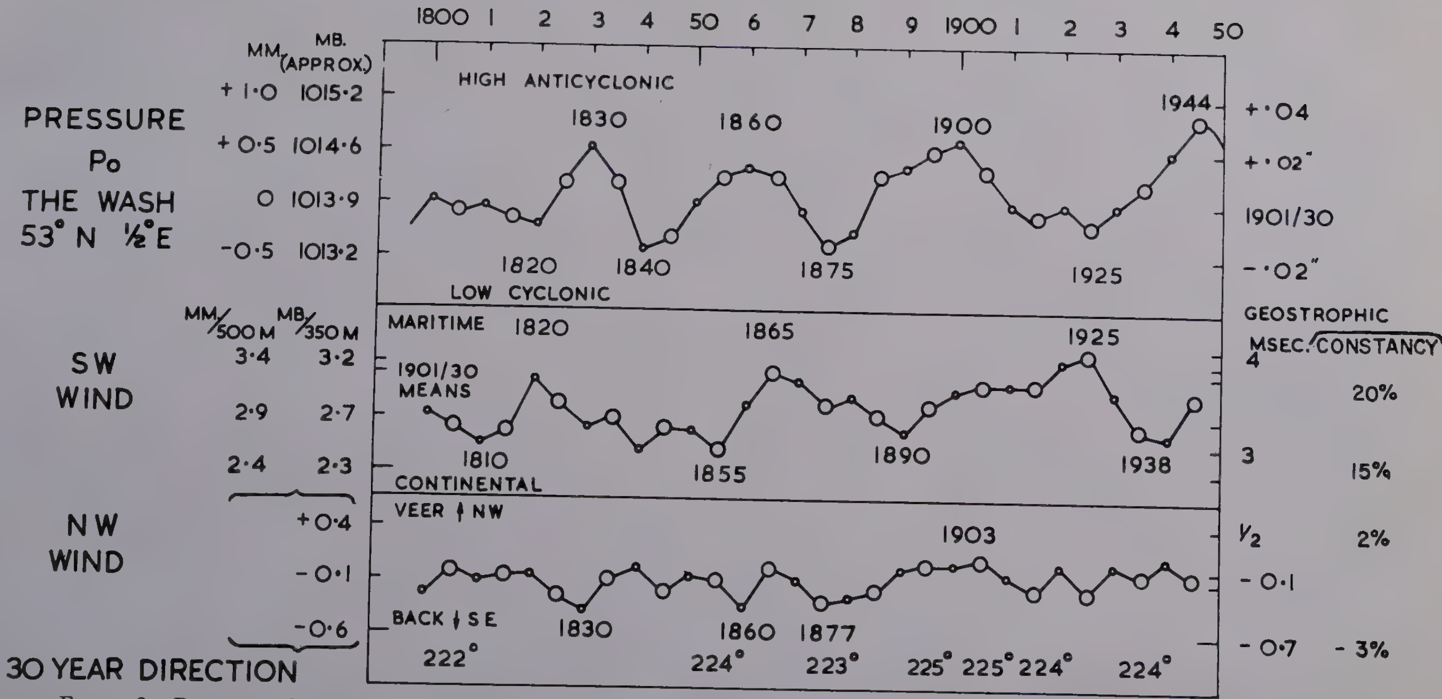


FIGURE 2. Pressure and pressure gradient, 1796 to 1950. Deviations of 10-year means from 1901/1930 normals. Reproduced from London Ph.D. thesis, 1958 (cf. Schove, 1961).

sure and temperature waves and, in the next few paragraphs, these waves in Europe and America will be briefly considered; also, some transatlantic parallels will be noted.

A mathematical investigation of cold and warm temperature waves in Europe since 1770 was made by Røstad (1955), who mapped the results. He did not notice that the direction of travel was generally north to south, and his dates are therefore all the more significant as evidence. Indeed the following mean

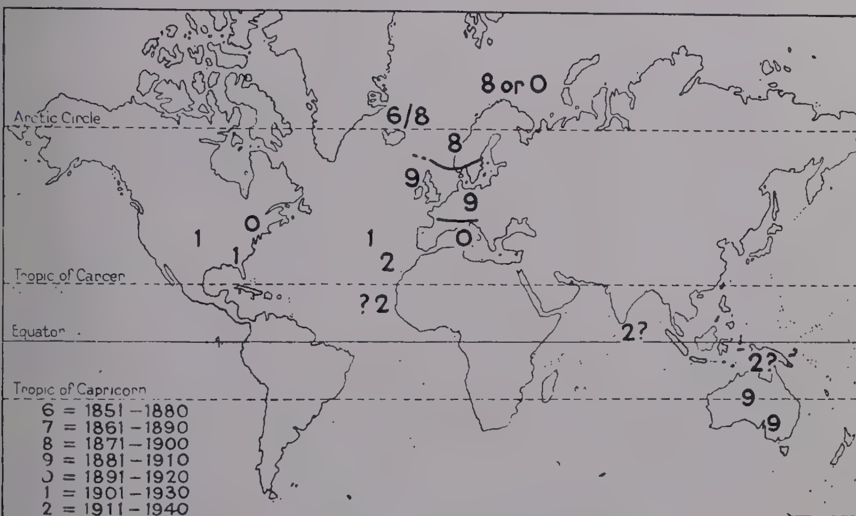


FIGURE 3. The P-max. time: that 30-year period in which average pressure was the highest since records began. Values south of 35° N should be redated to ca. *1890/1891.* The Pacific is probably ca. *1920/1923.* Reproduced by permission of the *Quarterly Journal of the Royal Meteorological Society* (Schove, 1950).

dates of temperature waves (relative to Copenhagen, Denmark) can be derived from his maps:

	Years
North Baltic (Haparanda)	−7
Middle Norway (Trondheim)	−5
Stockholm, Sweden	−5
Leningrad, U.S.S.R.; Poland (Vilna)	−4
Denmark and Scotland (Edinburgh)	0
Berlin, Germany; Prague, Czechoslovakia; and Vienna, Austria	+1
Paris, France; the Netherlands; Turin, Italy; and Geneva, Switzerland	+2
London, England	+3

Warm waves in Denmark thus occurred ca. 1824, ca. 1863, and ca. 1897, and cold waves ca. 1839 and ca. 1890.

Moving 10-year means of United States temperatures suggest that t–max. or warm waves (ca. 1825 to 1830, ca. 1849, ca. 1865, ca. 1897, ca. 1934, and ca. 1950) and t–min. or cold waves (ca. 1835 to 1840, ca. 1870, ca. 1888, ca. 1918

to 1921, and ca. 1938 to 1946) were almost simultaneous over the eastern half, although south steering systems did penetrate the United States (during Little Ice Age, III) from 1820 to 1850. Generally the American temperature waves

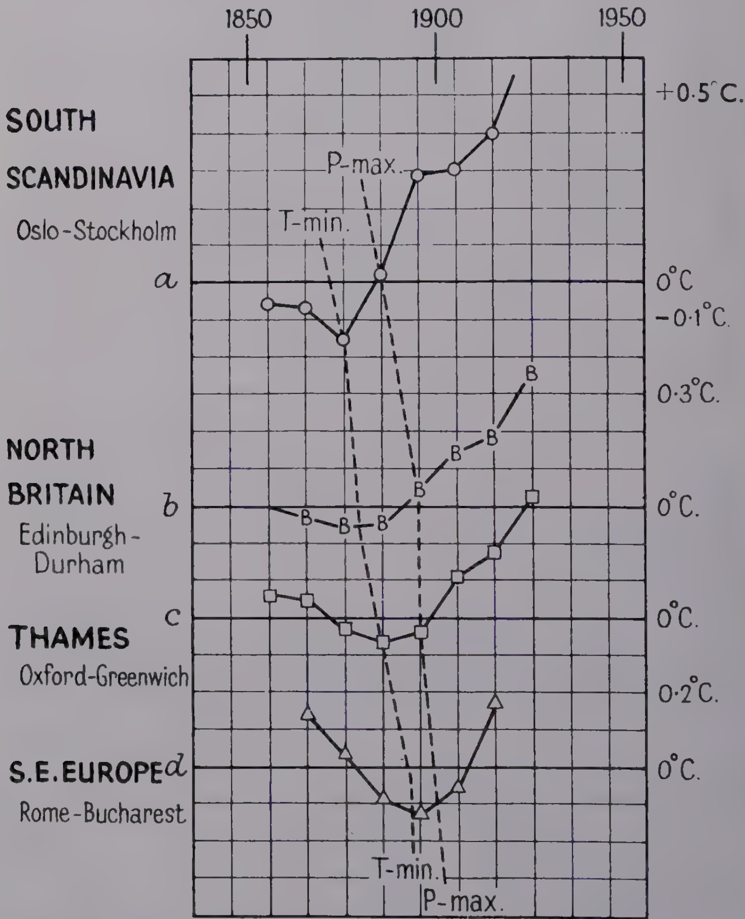


FIGURE 4. Climatic fluctuation in Europe and the Atlantic. Temperature-trend curves for Europe, north to south, relative to 1851/1900 normals. The P-max. followed the T-min. by about 10 years. Reproduced by permission of the *Quarterly Journal of the Royal Meteorological Society* (Schove, 1950). For pressure-trend curves, see Schove, 1950, Figure 2.

were in phase with northwest Europe, but in New England they averaged about one year earlier, and in South Carolina about one year later.

South steering in United States rainfall (April to August) is indicated in Flohn's diagram in this monograph. If r′ represents rainfall of the season April to August, the main r′–max. and r′–min. in the three areas are approximately as follows:

		First r′–max.	Main r′–min.	Last r′–max.
Middle West	A	ca. 1901	ca. 1934	ca. 1943
Prairies	B	ca. 1903	ca. 1936	ca. 1944
Mountains	C	ca. 1911	ca. 1937	ca. 1944

Considering the year as a whole the r–max. in the United States in the first period also traveled from north to south between ca. 1902 and ca. 1910, but there were further wet phases about 1915 and 1940 that did not show south steering. The main r–min. moved from north to south between ca. 1928 to

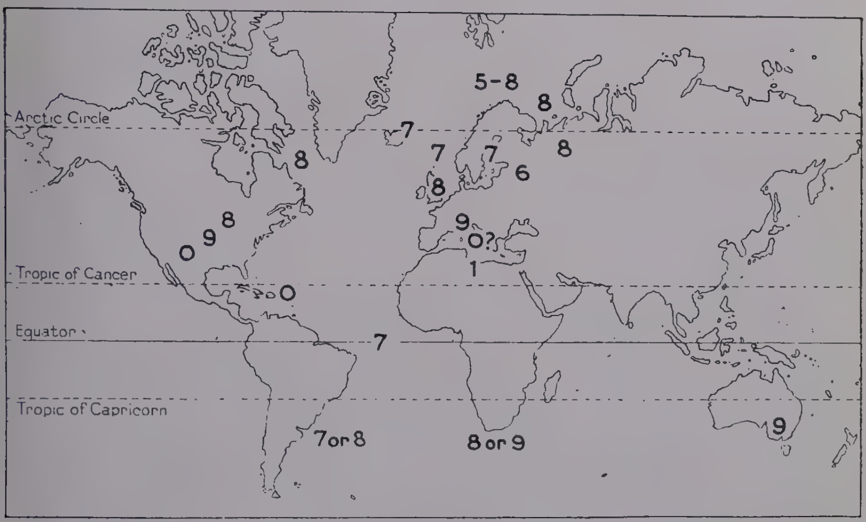


FIGURE 5. Climatic fluctuation in Europe and the Atlantic. The last T-min. time: that 30-year period (since 1851/1880) in which mean temperature was the lowest. Values south of 35° N should be redated. The tropical zone, using data from Callendar (1961) is thus 9, that is, ca. *1896* or 1882 to 1911. Reproduced by permission of the *Quarterly Journal of the Royal Meteorological Society* (Schove, 1950).

1932 and ca. 1934 (Tannehill, 1940, Figure 54). In this case as in many others the warm (t–max.) and dry (r–min.) phases apparently coincide, but as on the average the phases seem to have occurred about 2 years earlier in the tropics (warm, ca. 1876, 1897 to 1898, 1930, and 1939; cold, 1888, 1913 to 1920, using data from Callendar, 1961) than in the United States; it is clear that south steering of moving 10-year means cannot be relied upon in the subtropics.

There is some suggestion of south steering in 30-year moving means of tree-ring data (Schulman, 1956) of the United States, according to a preliminary analysis. The f(R) max. and f(R) min. in the Colorado area appear to be preceded by similar phases at Banff, Canada, and the Missouri area (by about 10 years), and probably the Snake area (by about 5 years), and they occur slightly later in southern California (0 to 5 years) and in the Rio Grande area

(by about 10 years). Moreover 15 years before an f(R) min. in the Colorado area there has often been much Icelandic ice combined with hot summers in Scandinavia and a warm phase in the United States itself. The significance of these relationships can be gauged better when they have been tested by computer analysis. In the meantime it may be mentioned that south steering in the North Atlantic is perhaps caused by a very slight rise of pressure following the upwelling of cold water with offshore winds.

Pressure systems that move south, as in FIGURE 3, appear to accelerate but also to become feeble south of 40° N. Other oscillations more important in the southern part of the United States will be considered later in these pages.

In 18th-century Europe the wind evidence is more reliable than the barometric, but the results in FIGURE 6 are consistent with the weather patterns. In the later 17th century, pressures are estimates based (Schove, 1961) on maps of weather patterns, and in some years winds have been estimated in the same way. Nevertheless the curve is still fairly reliable, and the general picture is consistent with Manley's (this monograph) temperature curves.

Before 1650 there are no pressure records and no long wind records, but the experience gained in dealing with the latter periods enables the documentary weather evidence to be used on its own. Maps of weather anomalies for each year and tables for each month in different parts of Europe have made it possible to determine the anomalous characteristics of different decades since at least 1090, and some of the results are included in TABLE 1.

In the early 17th century we have no barometric evidence, but there is a very clear alternation of anticyclonic and cyclonic decades in the period 1610 to 1660.

There is no indication in the historical record of another 20-year cycle of this kind that might otherwise seem to fit the double-sunspot cycle in tree rings and varves, found by some workers and denied by others in these pages.

The 16th-century climatic patterns may be studied in detail (Schove, to be published). Two distinct methods of estimating winter temperature lead to almost identical results: one is based on formulae, using the recorded warm (w) and cold (c) months (3w-c normally gives the most consistent results); the other on the amount of ice and of river floods caused by thawing (TABLE 1). The fall of winter temperature after 1540 (a winter, T–max. ca. *1529* to *1532*, followed by a winter, T–min., ca. *1559*) is in my view the first indication (Schove, 1949*a*; TABLE 2) of the onset of the Little Ice Age, whose phases determined originally from European documentary evidence fit so well the pattern found by Heusser (this monograph) both in Alaska and in southern South America.

Absolute temperature changes are not so easy to determine about 1300, although it is clear that this date is approximately that of a temperature recession both in Europe and in North America. In both continents grain crops were affected by the change to cooler summers (Smiley; Griffin; and Woodbury, all in these pages).

Cold phases in Europe probably correspond to cold phases in North America and can be noted at the following dates (Schove, 1958*b*): ca. *1140*, ca. *1220*, ca. *1315*, ca. *1445* and ca. *1605* to *1615*.

Warm phases occurred ca. *1185*, ca. *1255* to *1285*, ca. *1375*, ca. *1470*, and ca. *1520*. Probable dry phases in Europe are noted ca. *1255*, ca. *1465*, ca. *1545*, and ca. *1730*. Incidentally, tree growth both in Alaska and in the Southwest

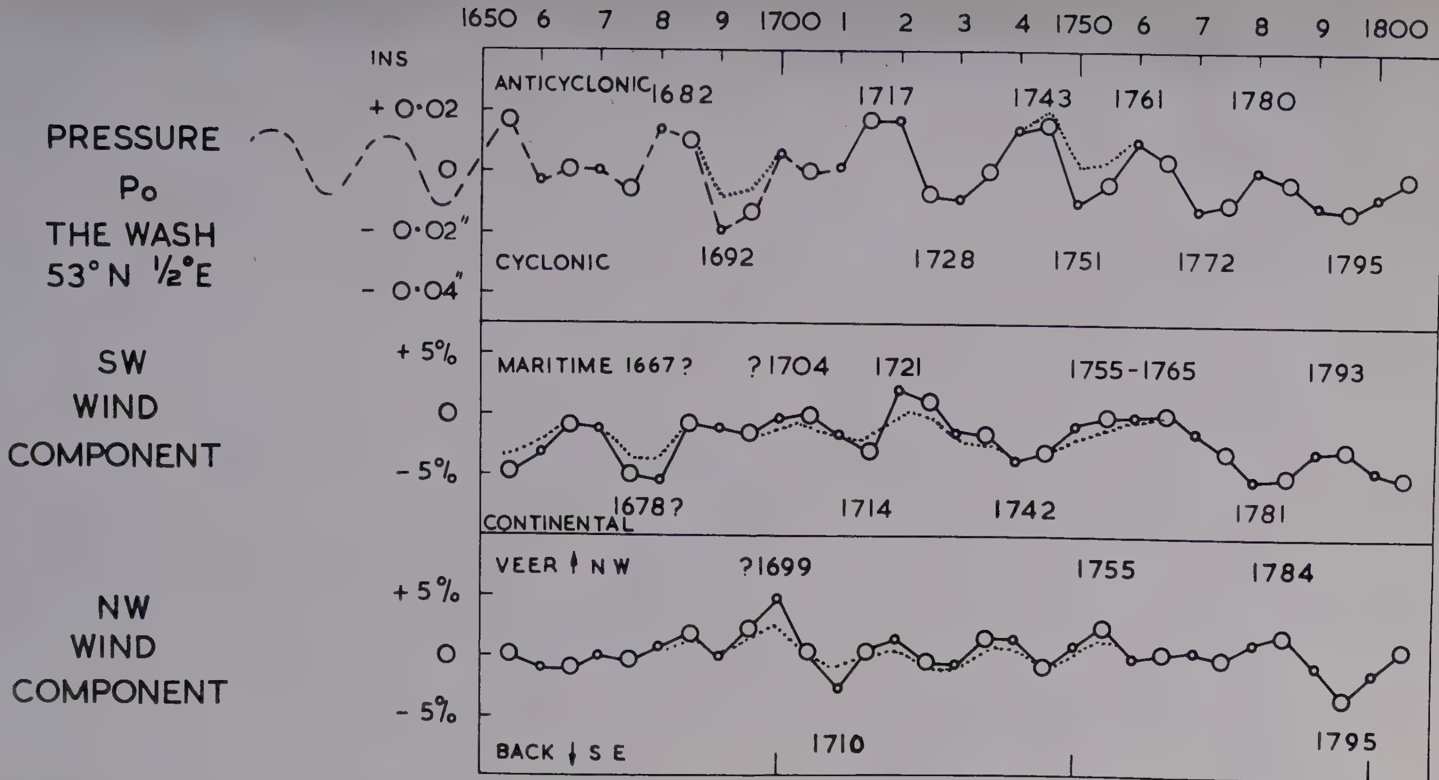


FIGURE 6. Pressure and wind components, 1651 to 1810. Means are provisional approximations to 1901/1930 values (cf. FIGURE 2). Dotted lines illustrate plausible alternatives based on the evidence of the weather patterns. Reproduced from London, Ph.D. thesis, 1958 (cf. Schove, 1961).

Table 1

Decades*	Auroral numbers†	American tree rings					Winter temperature, Europe (F)	Pressure, Europe (G)
1100s	50	Wet, Snake; dry, Colorado					(+5)	−1
1111/20	70	Wet both areas					(−2)	+1
1120	80	Still wet, Colorado					(−3)	−1
1130	80						+3	+3
1140	40	Long dry, Colorado, 1131 to 1161					−7	−3
1150s	50	Long dry, Southern California 1140 to 1190					+1	−1
1160	45	Dry Snake area 1157 to 1180					+1	0
1170	65						+2	+2
1180	60	Wet phase begins generally					+10	+2
1190	50	Wet					+6	−3
1200s	70	Wet					−2	+4
1210	50	Wet phase ending					−7	−1
1220	60	Dry, Colorado; Wet, Snake and Missouri					(+1)	−4
1230	30						(−1)	−3
1240	50	Dry phase spreads south					+6	+5
1250s	50	Droughts reach Rio Grande (Smiley, this monograph)					(+1)	−3
1260	45						+6	+5
1270	45						+4	−4
1280	40	Drought ends, Missouri					+9	+2
1290	30	Last decade of drought elsewhere					+3	+4
1300s	60	Long wet phase begins, Colorado					−1	0
1310	40						−3	−5
1320	45	Wet spreads to Snake, Mo., Southern California					+1	+1
1330	45	Wet everywhere					+1	+1
1340	30	Drier Colorado					+7	−3
		N. W.	S. W.	N. E.	S. E.	Mex.		
1350s	50	Drier Snake also.					0	+2
1360	60	dr‡	d	·	·		+5	−2
1370	80	·	r				+4	+1
1380	65	·	r		r		+2	+1
1390	40	·	·		r		−1	+1
1400s	40	d	d	d	r		+1	−1
1410	25	r	·	r	d		+5	+2
1420	25	·	·	·	·		+5	+1
1430	35	d	d	r	·		−6	−3
1440	40	r	dd	·	r		+1	+2

TABLE 1—*Continued*

Decades*	Auroral numbers†	American tree rings					Tree rings, Europe						
		N. W.	S. W.	N. E.	S. E.	Mex.	(A)	(B)	(C)	(D)	(E)	(F)	(G)
1450s	40	d	d	·	d					−4		−5	−1
1460	35	·	d	r	d					+1		−2	−1
1470	25	·	r	dr	d		−5			−2		+9	+3
1480	30	r	rr	dr	dd		+6			+4		+2	−2
1490	30	·	d	dd	d		−1			−5		−1	−3
1500s	35	·	·	dr	dr		−1	(−8)	−2	−4		+1	+1
1510	40	d	dr	rr	r		+2	−4	0	+2		+7	+2
1520	60	d	d	r	r		−3	0	+2	−1		−1	−5
1530	65	dr	·	dr	r		+4	+8	+5	+3	(+1)	+6	+6
1540	60	d	dr	dr	·		−5	−2	−12	−2	−1	+4	−3
1550	65	dr	rrd	r	r		−1	+4	+10	+5	0	−2	+2
1560	70	·	r	r	·		+3	−4	+6	+1	−2	+1	−1
1570	75	r	dd	d	d		−1	−2	−6	−3	0	−4	−1
1580	60	dr	dd	dd	d		0	−1	−8	−6	+3	−3	+3
1590	45	r	dr	r	d		−4	0	−3	−4	+2	−4	−1
1600s	50	r	rr	d	d		+7	−1	+10	+2	−2	+2	(0)
1610	50	rd	r	r	drr		−7	+6	+9	+5	0	+2	(+3)
1620	60	dd	dd	d	d		−1	−5	−9	−3	+2	−5	(−4)
1630	50	d	d	d	drr		+3	+5	−8	+2	−2	0	(+4)
1640	25	·	r	rd	r		−10	−1	+13	+3	0	+1	(−3)
1650	30	d	d	d	r		+9	−2	−4	+1	−5	0	(+3)
1660	25	r	d	d	rd		+1	−3	−8	−4	−3	−5	(−1)
1670	30	r	d	·	d		−7	+3	+9	+2	−3	−4	(−2)
1680	25	d	r	·	d		0	0	−3	−2	+3	−1	(+4)
1690	20	·	r	·	·		+4	+3	+3	+1	+4	−8	(−4)
1700s	25	r	·	·	·	d	+2	−4	−3	−3	+4	(+5)	(0)
1710	35	·	·	rd	·	dd	−9	−3	+1	0	+4	−2	+5
1720	60	·	·	·	rd	r	+4	+7	+4	+2	−4	+2	−4
1730	60	·	d	r	dd	r	+3	−2	−12	−6	−1	+4	−1
1740	40	·	rr	·	dr	·	−7	−3	+8	+4	+1	−4	+4
1750	36	rrd	d	d	·	dr	+6	+7	−5	0	−3	−2	−3
1760	57	·	·	r	r	·	−3	−7	+9	+1	0	−4	+2
1770	70	d	rdd	·	d	dd	+6	+3	−6	−3	+5	−2	−2
1780	72	dd	dr	rr	d	·	−9	−5	−7	0	+3	−7	+2
1790	28	dd	rd	d	r	r	+7	+3	+6	+3	0	−4	−2
1800s	26	·	rd	r	rdd	·	−1	+3	−2	−2	0	−4	+2
1810	22	r	d	·	·	rr	+7	−3	+3	0	−4	−4	−2
1820	33	·	ddr	dr	r	dd	+11	+3	−10	0	+4	+6	+2
1830	67	r	r	r	rr	r	−4	−6	+18	+7	−3	−2	+2
1840	57	dd	d	rrd	·	r	+1	+1	−16	−1	−2	+2	−5
1850	46	·	·	d	d	·	+1	+3	+5	−1	−1	−4	+3
1860	53	r	d	dr	dr	·	−6	−1	+7	−1	+3	+4	+3
1870	40	dr	·	dr	dd	·	+4	−1	−9	−2	0	0	−5
1880	35	rrd	dr	rd	d	r	+1	+4	+5	−1	0	0	+2
1890	45	ddr	dd	dr	dd	d	+3	−5	−7	−2	+1	+2	+2
1900s	37	rr	rr	dr	·	·	−7	+7	+4	+1	−3+	+4	+1
1910	41	rd	rr	rdd	rr	r	+5	−8	+5	+7	−2+	+9	−1
1920	42	dd	r	·	r	·	(−)	+3	−2	−3	−2+	+6	−2
1930	54	·	rd	d	·	rdd	(+)	( )	(−3)	(0)	( )	+9	−1
1940	74	( )	d	d	( )	( )	( )	( )	( )	( )	( )	(+)	(+)

* Decades are of the form 1101–10.

† Auroral numbers to 1700 are intended to correspond to sunspot numbers. Thereafter, sunspot numbers are as in Schove (1955), with approximately 10 per cent increase to 1740.

(Footnotes concluded on p. 118.)

TABLE 1—*Concluded*

‡ Droughts and moisture as reflected in the semiarid zone, from unpublished maps largely based on data in Schulman (1956). The north-south division is the Rockies. The NW is north of 42° N, but the NE is north of 39° N. The letters dr mean dry then wet. Mex. signifies west central Mexico.

§ Columns (*A*) through (*C*) are keyed as follows:

(*A*) North Scandinavian trees (adapted from Schove, 1954*a*). Relative anomalies. Selected to indicate summer temperature.

(*B*) Alaskan trees (adapted from Schove, 1953*e*). Relative anomalies. Selected to indicate summer temperature.

(*C*) Western United States trees (adapted from Schove, 1953*e*, columns 3 and 4). Relative anomalies.

(*D*) Southern United States trees (Schulman, 1956, Tables 78, 70, and 50, and standardized curves, pp. 86, 87).

(*E*) Java minus "Pacific." Differences between tree growth in Java and Argentine-plus-west central Mexico. Unreliable at present as a measure of pressure parameter. Values from 1900 affected also by disease in Argentine trees.

(*F*) Winter temperature in northern Europe (45° to 60° N). Based mainly on Schove, 1953*e*, but with revisions. Estimated from ice data and documentary evidence. Winter in this column is one half a year, for example, November to April, and latitude is 45°/55° N to 1700, 53° N after 1700. From 1700, Manley's (1957) temperature series (November to April) (C × 10) and length in quarter days of Baltic ice-free season at Leningrad, Union of Soviet Socialist Republics (Sokolov, 1955), have been given equal weight.

(*G*) Pressure in northwest Europe (for example, 50° N 5 E) estimated from documentary data to 1650. Relative anomalies. Adapted from Schove, 1953*e*, with revisions. Pressure to 1600: 45°–55° N, based on documentary evidence of dryness and similar factors. After 1600: "The Wash," 53° N ½° E (Schove, 1953*e*, with revisions, and 1958).

TABLE 2*

PATTERN OF THE LITTLE ICE AGE IN EUROPE (40 TO 55° N) AS SUGGESTED BY THIRTY-YEAR MEANS OF DOCUMENTARY DATA

Period	Glaciers	Weather type	Winters	Summers
1451 to 1540	Preglacial	Very maritime	Very mild	Becoming cool
1541 to 1600	Little Ice Age, Phase I	1a Continental	Cold	Hot at first
(1590) 1591 to 1650		1b Moist, cool	Cold at first	Cool
1651 to 1680		1c Very continental	Very cold	Hot
1681 to 1740	Interglacial	Maritime	Mild	—
1741 to 1770	Little Ice Age, Phase II	Continental	Cold	—
1771 to 1800	Lull	Continental	Cold	Hot
1801 to 1890	Little Ice Age, Phase III	Continental	Cold	—
1891 to 1950	Interglacial	Very maritime	Very mild	—

* Thermometric series prepared for central England by Manley (1957) now make it possible to study the above pattern over the last 250 years in more detail (reproduced from the *Quarterly Journal of the Royal Meteorological Society*, 1949, p. 176). Summer and winter temperature anomalies are not in general opposed to each other, but if the differences between the two are termed continentality, for example c, the main c–max. is dated ca. 1779 and the main c–min. ca. 1924. In Vienna, Austria, the change from continental to maritime took place between ca. *1885* and ca. *1912* (see Steinhauser, 1960).

is greater in decades that in Europe are relatively warm, maritime, and anticyclonic.

There are several reasons for parallelism between minor transatlantic fluctuations: both regions are affected by the main pressure "center of action" off west Greenland (see below), and there is probably a parallelism between the pressure in Northwest Europe and the pressure off western Canada. Unfortunately

the lack of any reliable pressure data in North America before 1873 (the Russian record at Sitka was not continued, and the only pressure series going back to 1841 is that for Toronto, Canada) makes it difficult to attempt a chart for North America corresponding to our FIGURE 2, although similar methods are being used (see below).

In concluding this survey the lack of any simple relationship between the European data and the sunspot data (TABLE 1) must be emphasized.

Sunspots and Climatic Fluctuations

Sunspots and climatic fluctuations may often run in parallel but, as will be explained (Schove, elsewhere in this monograph), the 11-year sunspot cycle is

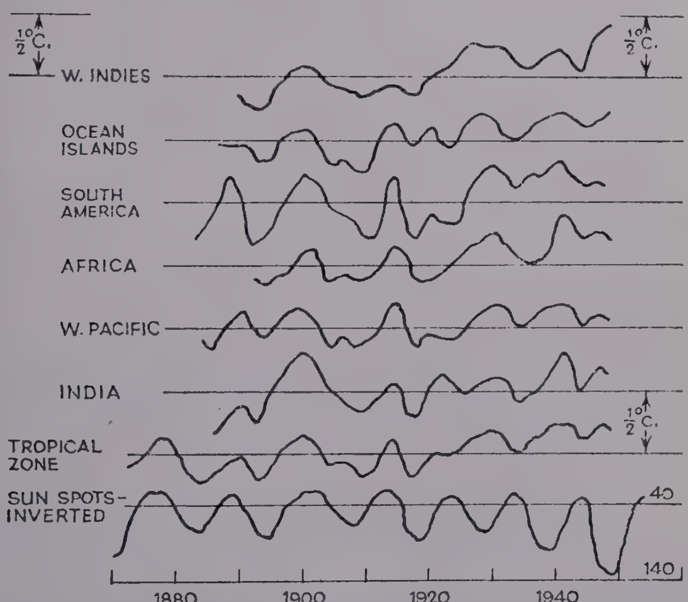


FIGURE 7. Temperature fluctuations in tropical regions, 5-year annual mean departures from mean 1901 to 1930. Reproduced by permission of the *Quarterly Journal of the Royal Meteorological Society* (Callendar, 1961).

often only an indirect cause. An instance of this is the well-known inverse relationship between sunspots and tropical temperatures. A world survey of temperature fluctuations has just been published (Callendar, 1961). The results are consistent with those found by Mitchell (elsewhere in these pages), but the two surveys illustrate different aspects. FIGURE 7 reproduces part of one of Callendar's diagrams, according to which the usual sunspot relationship was evidently reversed in the period 1921 to 1950. A pressure parameter is determined (Schove, elsewhere in this monograph) that may reflect more accurately than sunspot numbers the effective solar radiation.

The significance of the variation of decadal means (or, for example, 11-year overlapping means) of auroral numbers (TABLE 1) from cycle to cycle may be mentioned here: the cold waves in the tropics and in the United States often

correspond to active solar cycles, although the present snowy winter of 1961 could hardly have been predicted from the great sunspot activity of the period 1955 to 1960.

The Spectrum of Early Time

The Spectrum of Time technique is being applied both to geological periods and to the Pleistocene.

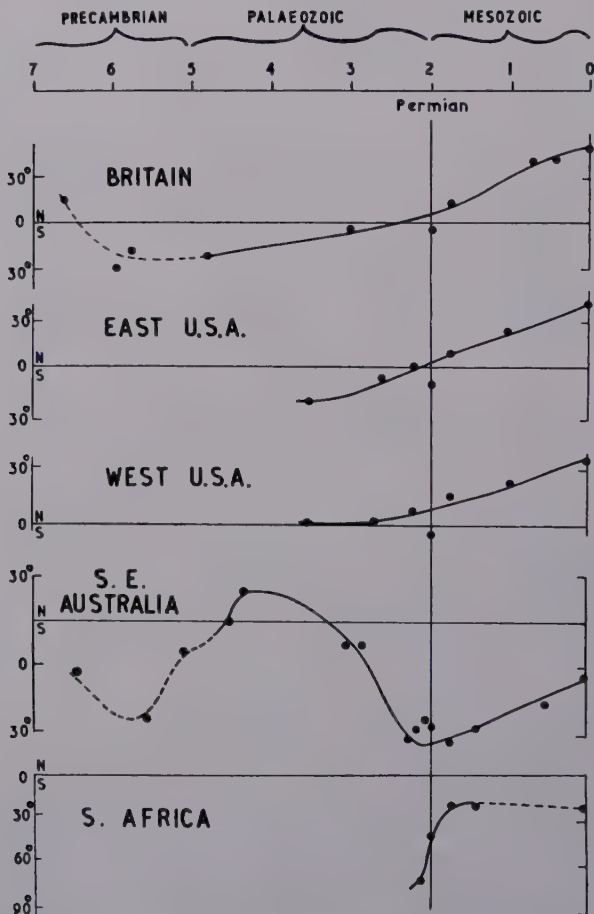


FIGURE 8. Variations in geomagnetic latitude in different continents. Reproduced by permission of *Geografiska Annaler* (Schove *et al.*, 1958, p. 219, adapted from Irving, 1957, 1958).

In the Permocarboniferous, for instance, it is possible to suppose that the glacials of the present southern hemisphere correspond with the pluvials and low sea levels of the cold phases of the present northern hemisphere. In a prior work (Schove *et al.*, 1958*a*) we did not actually make this assumption, but some of our results are relevant to the subject matter of this monograph and are reproduced because of their interest. Distribution maps of different fossils of

datable horizons are being collected and compared with paleomagnetic evidence for latitude and paleowind evidence for pressure patterns.

The change of latitude with time for the continents, according to the paleomagnetic evidence, is shown in FIGURE 8. The map of winter temperatures in

FIGURE 9. The climatic geography of the Permian. Winter sea-surface temperature in the Permian as indicated by the criteria used by Stehli (1957, Figures 1 to 5), for the areas where outcropping marine rocks are available. "Warm" fauna on each of Stehli's maps have constituted 2 points, so that places with warm-type brachiopods and foraminifera on all 4 maps received 8 points. Such places were presumably equatorial and are indicated in the above graph by squares. Reproduced by permission of *Geografiska Annaler*. (Schove *et al.*, 1958, p. 225).

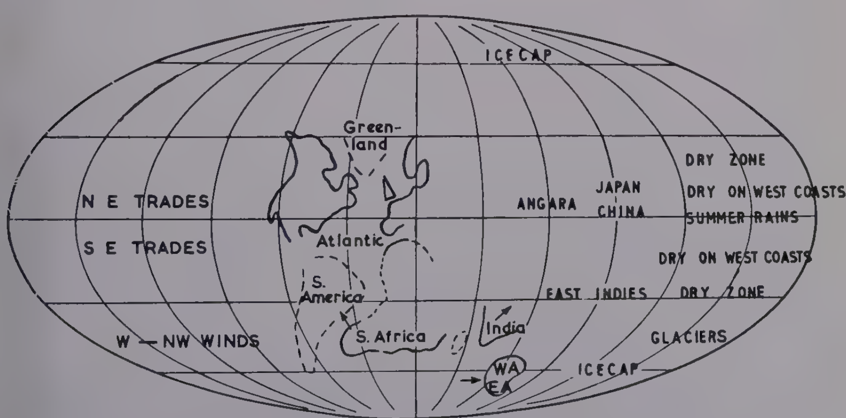


FIGURE 10. The latitudes and orientations of the continents in the Permian. Reproduced by permission of *Geografiska Annaler* (Schove *et al.*, 1958, p. 221).

the Permian, derived from the work of Stehli (1957) who believed that the continents had not moved, is shown in FIGURE 9. Finally, the relative positions of the continents as shown in FIGURE 10 constitute an interpretation that accords best with the three types of evidence.

In the Pleistocene the detailed patterns of the interglacials found by Emiliani in deep sea deposits will soon be found to match the detailed patterns of the interglacials as found by pollen analysis. In so far as the last glacial is concerned, the spectrum of the last 80,000 years of time is very neatly shown in the diagram of Flint and Brandtner.* The key dates are of global significance. Radiocarbon dates at significant pollen horizons show clusters at specific B.P. dates in different parts of the world. The indication of hot summers noted in that diagram ca. 60,000 to 55,000 B.P. accords with that noted by Solecki (elsewhere in these pages) from date stones found in mountains of northern Iraq; it also accords with the hot summers implied by spruce cones found in the Cheshire Plain of England and discussed recently by Simpson.

One postglacial date in the Spectrum of Time will be mentioned: 2750 ± 100 B.C. (4700 B.P.) according to current radiocarbon dating or, roughly, 3000 B.C., according to absolute solar chronology. These at least were the dates announced at the Union Géographique Internationale (U.G.I.) meeting at Stockholm, Sweden (Schove, 1960), as representing the time when the dryness of the subtropical anticyclones extended both into central North America and Northwest Europe.

Very few solar cycles before 200 B.C. can be dated with any confidence, but some suggestions will be made in my article "Tree Rings and Climatic Chronology" (this monograph) to indicate regions where the effects of subsequent solar cycles, as dated in Schove (1955*b*), might be investigated in future studies of varves and tree rings.

References

CALLENDAR, G. S. 1961. Temperature fluctuations and trends over the earth. Quart. J. Roy. Meteorol. Soc. **87:** 1–12.

DEWEY, E. R. 1958. The length of the sunspot cycles. J. Cycle Research. **7**(3).

DEWEY, E. R. 1960. The 200 year cycle in the length of the sunspot cycle. J. Cycl. Res. **9** (2).

GLEISSBERG, W. 1958. J. Brit. Astron. Assn. London, England.

MANLEY, G. 1957. Climatic fluctuations and fuel requirements. Scott. Geogr. Mag. **73:** 19–28.

SCHOVE, D. J. 1947. The sunspot cycle before 1750. Terrestrial magnetism. **52:** 233–238. Williams & Wilkins. Baltimore, Md.

SCHOVE, D. J. 1948*a*. Sunspot epochs, A.D. 188–1610. Popular Astronomy. **56:** 247–262.

SCHOVE, D. J. 1949*a*. European raininess and European temperatures, A.D. 1500–1950. Quart. J. Roy. Meteorol. Soc. **75:** 175–179; 181.

SCHOVE, D. J. 1950*a*. The climatic fluctuation since A.D. 1850 in Europe and the Atlantic. Quart. J. Roy. Meteorol. Soc. **76**(328): 147–165. *Cf.* Meteorol. Abst. & Bibl., **1950:** 362 and Geogr. Rev., 1951. **41:** 656–660.

SCHOVE, D. J. 1950*d*. Visions in north-west Europe and dated auroral displays (A.D. 400–600). J. Brit. Arch. Assoc. 3rd ser. **13:** 34–49.

SCHOVE, D. J. 1951*c*. The earliest dated sunspot. J. Brit. Astron. Assoc. **61:** 22–24.

SCHOVE, D. J. 1951*e*. Sunspots, aurorae and blood rain. Isis. **42:** 33–138.

SCHOVE, D. J. 1953*e*. M.Sc. Thesis. London, England.

SCHOVE, D. J. 1953*f*. The preliminary reduction of early barometric and wind data. Paper 10, in U.G.G.I. Publication AIM No. 9/c. Proces-verbaux des seances de l'Association de Meteorologie, Memoirs and Discussions. : 187–193.

SCHOVE, D. J. 1954*a*. Summer temperatures and tree-rings in north-Scandinavia, A.D. 1461–1950. Geogr. Ann. Stockholm. **36**(H 1–2): 40–80.

SCHOVE, D. J. 1955*b*. The sunspot cycle, 649 BC–AD 2000. J. Brit. Astron. Assoc. **60** (2): 139.

* *Weather*. 1961. June/July. Vol. 15.

SCHOVE, D. J. 1956*a*. Sunspot maxima since 649 B.C. J. Brit. Astron. Assoc. **66**(2): 59–61.
SCHOVE, D. J. 1958. Ph.D. Thesis. London, England.
SCHOVE, D. J. 1960. Climatic chronology and historical geography. U.G.I. International Geog. Congr. Stockholm. Abstr. 256.
SCHOVE, D. J. 1961. Wind and pressure in N.W. Europe since 1630. Geogr. Ann. Stockholm, Sweden.
SCHOVE, D. J. & A. W. G. LOWTHER. 1957. Tree-rings and medieval archaeology. Med. Archaeol.
SCHOVE, D. J. & P. Y. HO. 1958. Chinese aurorae, AD 1048–AD 1070. J. Brit. Astron. Assoc. **69**: 295–304.
SCHOVE, D. J., A. E. M. NAIRN & N. D. OPDYKE. 1958*a*. The climatic geography of the Permian. Geogr. Ann. 3–4.
SCHULMAN, E. 1956. Dendroclimatic Changes in Semi-Arid America. Univ. Ariz. Press. Tucson, Ariz.
SIMPSON, I. M. 1959. Quart. J. Geol. Soc. **115**(107–122): 117. London, England.
SOKOLOV, S. S. 1955. Trans. by E. R. Hope from Priroda. 7: 96–98. Defence. Scient. Inform. Service, DRB, Canada.
STEHLI, F. G. 1957. Am. J. Sci. 255.
STEINHAUSER, F. 1960. Geof. e Met. Genova, 111: 112.
TANNEHILL, I. R. 1947. Drought. Princeton Univ. Press. Princeton, N. J.
VISSER, S. W. 1959. Med. en Verh. **75**. (Kon. Ned. Med. Inst.)
WATANABE, N. 1959. The direction of Remanent Magnetism of baked earth . . . in Japan. J. Faculty of Science. Univ. Tokyo. **5**(Vol. 11): 1–188.

CLIMATIC CHANGE WITHIN HISTORICAL TIME AS SEEN IN CIRCULATION MAPS AND DIAGRAMS

H. H. Lamb
Meteorological Office, London, England

Introductory

Solar radiation supplies the energy, and the terrestrial atmospheric circulation is the mechanism of climate. Grading of the radiation budget by latitude explains the basic succession of climatic zones from equator to pole, and gives this sequence the stamp of permanency, although the positions of the zones may undergo some general poleward or equatorward shift from time to time or epoch to epoch. It is the atmospheric circulation that produces the weather and explains the distribution of such events as orographic rainfall, snow and snowfields, thermal instability, thunderstorms and tropical hurricanes, the extent of oceanic, continental, and arid climates. Thus it is to be expected that circulation maps and diagrams for different epochs should be a valuable tool for clarifying our understanding of climatic shifts and changes.

The scientific purpose of historical studies of climatic change is to reveal the behavior of the atmospheric circulation and throw light upon the essential controls, external and internal, that may be expected to operate in the future as in the past. This paper presents some of the first results of studies in the British Meteorological Office with these aims in view.

Investigational Method: Example of Mean Circulation Maps and Their Interpretation

To study climate and the longer-term changes of the atmospheric circulation, the observational material must be condensed in some way. It has been found convenient to work chiefly with 5-day means and monthly means, sometimes averaged over periods of years. Monthly data are by far the most abundant over the period since measurements by meteorological instruments began, and appear to provide a sufficient understanding of the seasonal round of the year for most purposes; 5-day or "pentad" averages may be used to reveal features of a finer seasonal structure of the year, which recur in most years and are likely to be glossed over by monthly means, although the details have little relevance to long-term climatic changes. Three-month or "seasonal" averages must sometimes be resorted to for early epochs where data are scanty but can probably be meaningful only in the extreme seasons of summer and winter when the trend curves of radiation, temperature, and circulation intensity are near their maxima or minima (first derivatives approximately zero). Annual averages of the atmospheric circulation pattern are generally to be avoided because the changes of circulation character between summer and winter, especially in the northern hemisphere, are too great to be summarized in this way.

Dynamic studies reveal that eddy transports may be as important as the transports due to the mean flow. It will be necessary therefore to acquire some idea of the probable frequency of cyclones and anticyclones and of various types

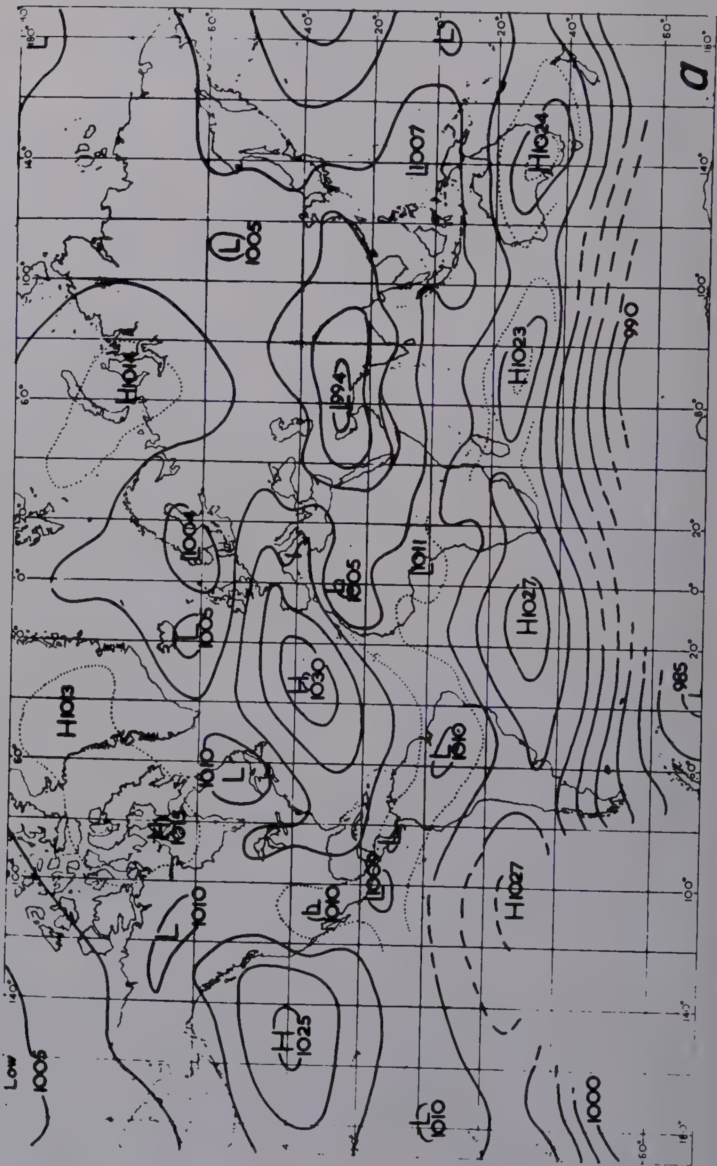
of instantaneous flow with different mean circulation patterns. Similarity between the prevailing features of the instantaneous pattern sequences and the main features of the period-average map is readily seen in the case of 5-day means, and also with the great majority of monthly mean patterns, but must become increasingly difficult (if not impossible) to elucidate with averages over longer and longer periods of the year.

World maps of the monthly values of atmospheric pressure at m.s.l., surface-temperature anomaly, precipitation anomaly, thickness (mean temperature) of the 1000 to 500 mb. layer and of 500 mb. height (topography) and the anomalies of these, have been analyzed regularly each month in the Meteorological Office at Harrow, England, for several years past, using the world-wide exchange of broadcast messages (in CLIMAT and CLIMAT TEMP codes) organized by the World Meteorological Organization (Lamb, 1957; Lamb and Johnson, 1960). The object of studying these charts has been to discover how the large-scale atmospheric circulation behaves from month to month (the regular seasonal changes and the succession of more or less anomalous patterns occurring in each particular year) and to understand something of how the more striking weather anomalies reported arise.

Experience soon showed that the monthly mean circulation maps at the surface and aloft are admirable for interpreting the significant weather features of the month and may be looked upon as a fundamental aid in climatology. Their use in following long-term changes and the development of spells of weather that determine the character of particular seasons is analogous to the role of the synoptic weather map in meteorology for interpreting day-by-day changes. This is because the prevailing positions of the main warm and cold air streams and the depression tracks and zones of anticyclonic subsidence so largely determine the weather at the places affected.

Average pressure and wind patterns for January and July printed in physical atlases have long been used for teaching the distribution of world climates. It has only recently become possible for the first time for the patterns over the whole world to be computed for a period of years that is strictly uniform everywhere. However, the same years *must* be used everywhere if all the climatic phenomena of the epoch are to be physically explained. The reason is that as soon as we begin examining individual months, years, and decades separately, we realize that climatic changes of greater or lesser degree are always going on.

The study of individual months is highly informative, as the examples in FIGURES 1, 2, and 3 clearly show. FIGURES 1*a* and *b* depict side by side the mean circulation patterns (at the surface) of two very dissimilar Julys in recent years; namely 1954 and 1955, which were respectively wet and fine in both northwest Europe and much of northeastern North America, with general reversal of the longitudinal placings (phase change) of the main pattern features in the higher latitudes of the North Atlantic and neighboring sectors. FIGURES 2a and b likewise compare dissimilar Januarys, those of 1940 and 1957, an example each of weak and strong circulations over the North Atlantic with prevailing easterly and westerly winds giving respectively severe and mild winter weather in western Europe; the differences over western North America


a
Low
1005
H1025
H1030
H1027
H1024
H1023
H1014
H1013
L1005
L1004
L1007
L1011
L994
990
985
1000
1010

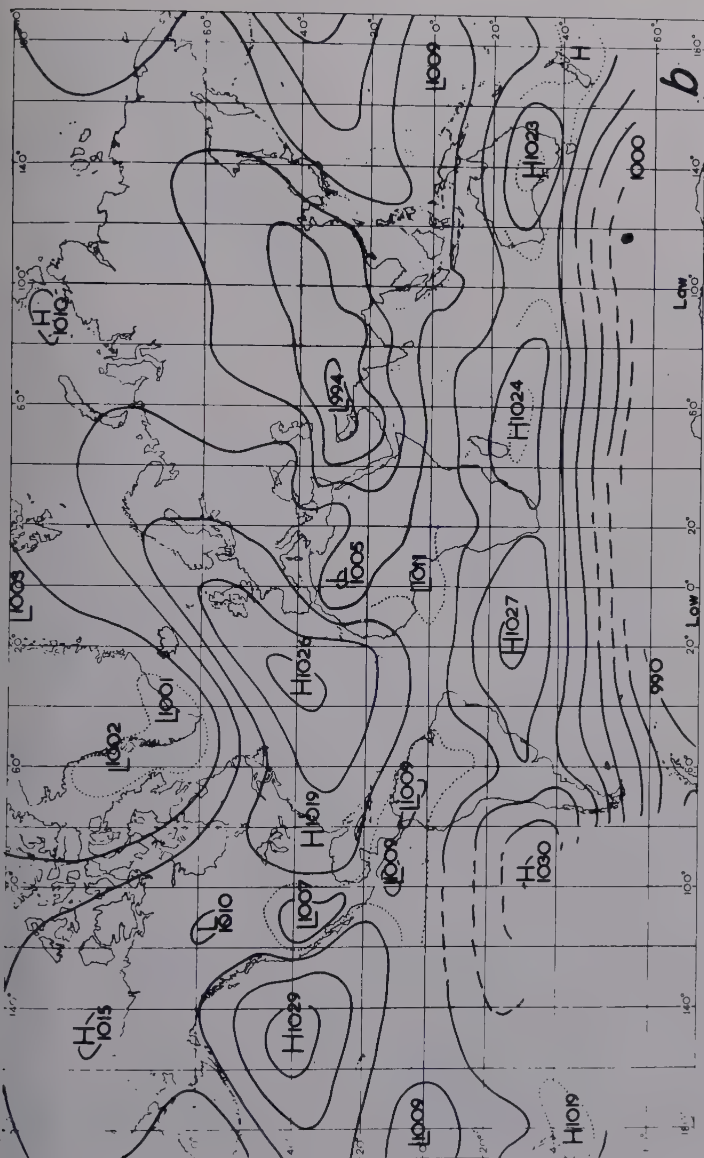
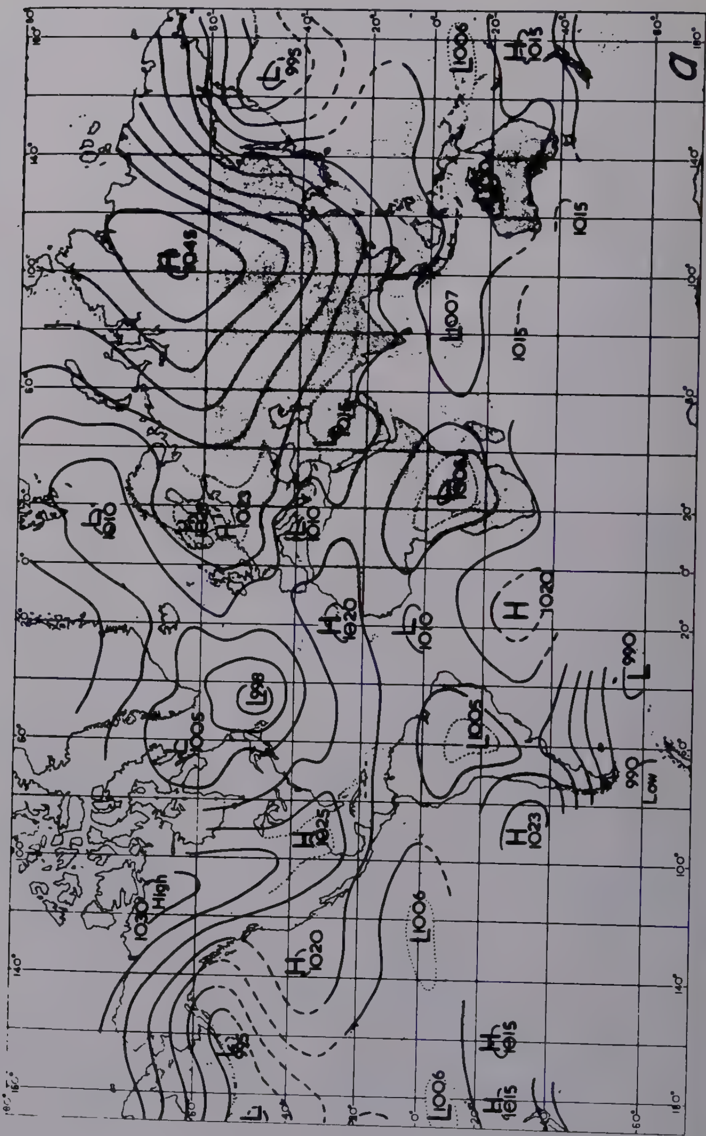


FIGURE 1. Monthly mean sea-level pressure over the world (*a*) July, 1954, and (*b*) July, 1955. Graph *a* shows a weak (graph *b* a strong) summer circulation in the higher northern latitudes, with great differences of pressure pattern between the two Julys in most of the northern temperate and arctic zones. Reproduced by permission of *Geografiska Annaler*, Stockholm, Sweden.


a
1045
995
1006
1013
1015
1007
1015
1010
1015
1020
1010
1020
1010
998
1005
990
990
Low
1005
1023
1025
1030
High
1020
1006
1015
1006
1015

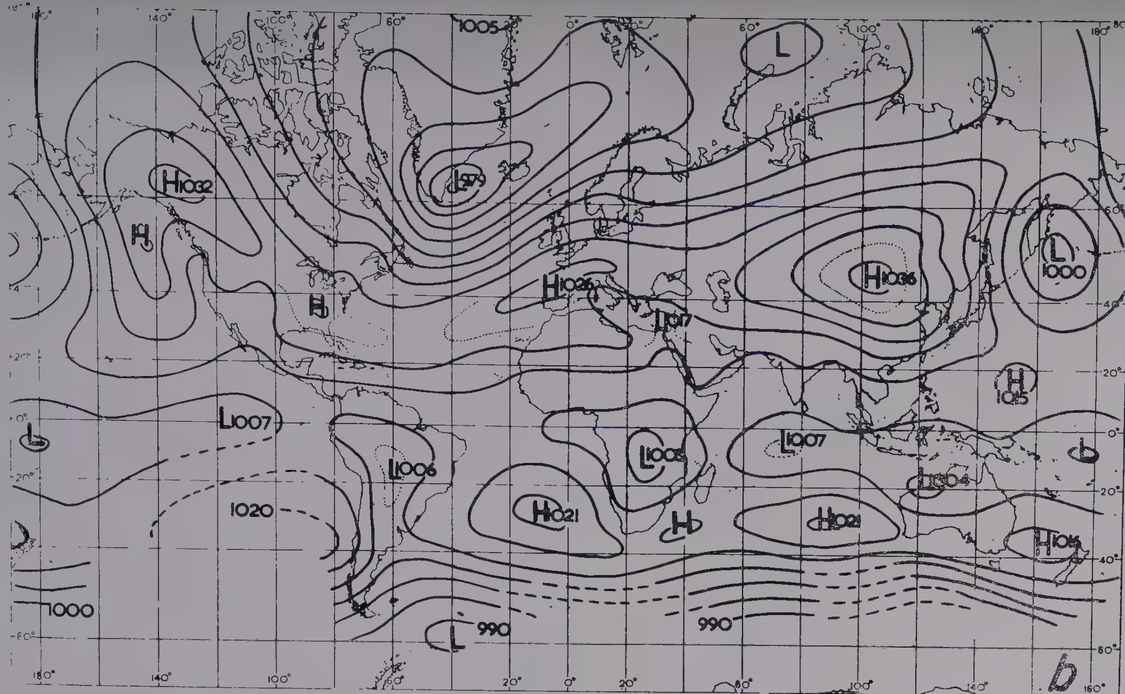


FIGURE 2. Monthly mean sea-level pressure over the world (*a*) January, 1940, and (*b*) January, 1957. Graph *a* shows a weak (graph *b* a strong) winter circulation in the zone of northern hemisphere westerlies. In 1940 the westerlies were so weakly developed over the ocean that easterlies developed over the continents in the temperate zone and covered even the western coasts, whereas in 1957 exceptionally strong westerlies developed over the Atlantic (but not the Pacific) and spread across most of the downstream continent. Reproduced by permission of *Geografiska Annaler*, Stockholm, Sweden.

are also noteworthy and in this case almost inverse to those over Europe. January 1959 (not illustrated) revealed another type of weak zonal circulation in which northerly outbreaks in the Norwegian Sea were a prominent feature.

FIGURES 3a and b show the mean surface-pressure pattern of October 1960 over the northern hemisphere compared with the so-called "normal," or (better named) period average, for that month over the years 1899 to 1938. This

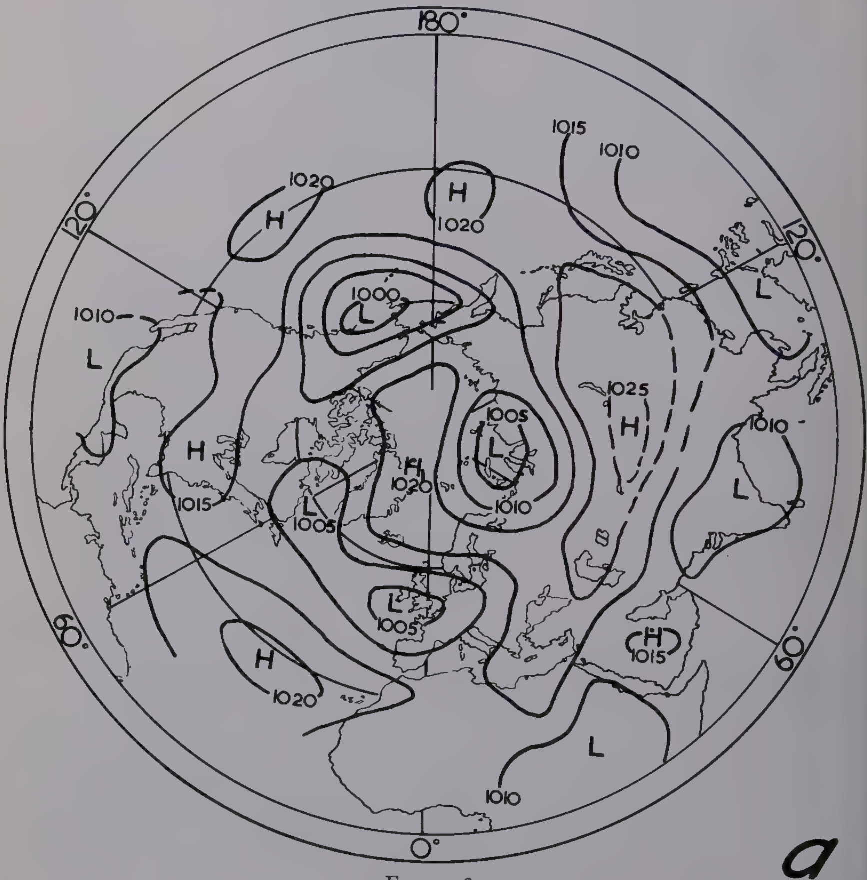


FIGURE 3.

example shows how important it is in these studies to distinguish anomalies both of circulation intensity and position. In October 1960 the "Iceland low" had about the usual central pressure, but was displaced over 1000 miles to a position covering southwest England, where 3 to 4 times the usual rainfall brought disastrous flooding. The prevailing depression track across the North Atlantic (low pressure trough on the mean map) was over 10 degrees of latitude south of normal in October 1960, and the entire circumpolar arrangement of the hemisphere pressure pattern (including both the polar "high" and the depression track corresponding to the main circumpolar westerly wind

vortex in the upper troposphere) appears "tilted" or displaced by about 10 degrees toward the Atlantic sector. This type of eccentricity of the circumpolar vortex is easier to detect when it occurs in the southern hemisphere with its simpler, zonal type of circulation, but is often suspected in the northern hemisphere too. October 1960 was an extreme case of a kind apparently unmatched

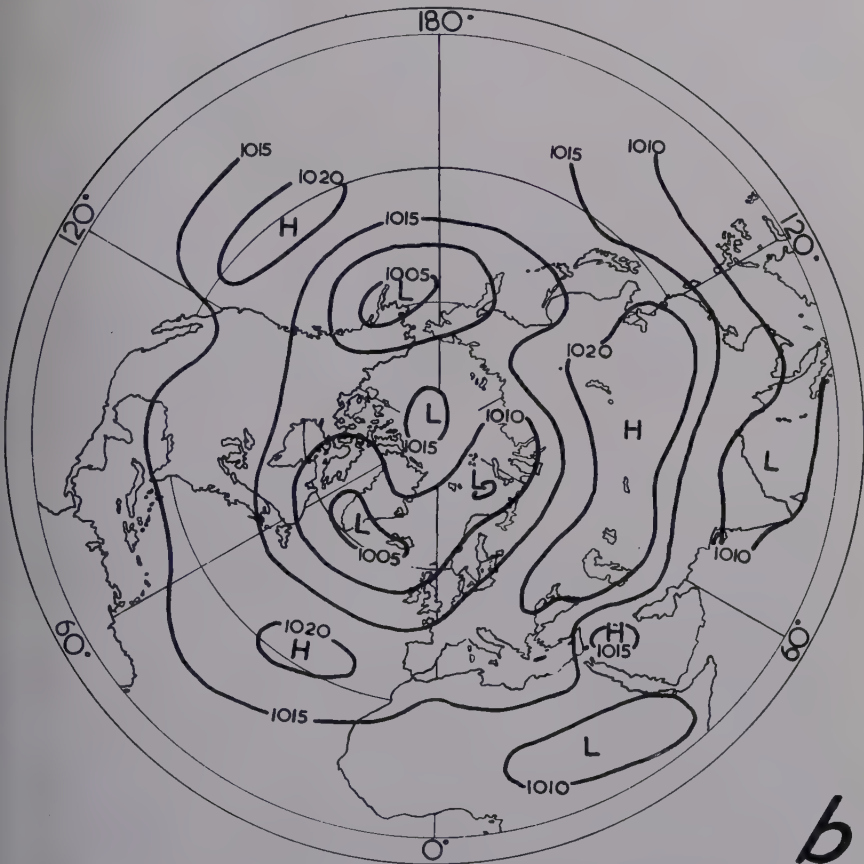


FIGURE 3. Monthly mean sea-level pressure over the northern hemisphere (map *a*) October 1960, and (*b*) the average for October 1899 to 1938. Map *a* shows an exceptional southward displacement of the depression track and belt of westerlies in the Atlantic sector and some shift of the entire circumpolar circulation toward the Atlantic sector. Reproduced by permission of *Geografiska Annaler*, Stockholm, Sweden.

in the 200 to 250 Octobers for which meteorological observations can be consulted. A similar although smaller equatorward displacement of the North Atlantic depression track however has been a prominent and often persistent characteristic of many months and seasons in the last few decades.

Identification of such trends in the mean circulation—namely (1) changes of intensity, (2) latitude shifts, and (3) changes of longitudinal spacing—facilitated by maps of this kind is an essential first step before either dynamic or physical

explanations can be arrived at. Comparisons of northern and southern hemispheres show that latitude shifts in the two hemispheres are sometimes toward or away from each other but sometimes in the same sense; that is, in any given sector northward or southward in both hemispheres. The maps are also of considerable interest to oceanographers and others, because they throw light, for instance, on changes of strength and position of the great ocean currents, changes in the drift (as well as thickness and extent) of the ice on the polar seas, anomalous migrations of fish and birds, and the smaller organisms they feed on. Similar monthly mean flow charts, more particularly for the higher levels in the stratosphere, are also of value in understanding the transport of ozone, volcanic dust, and nuclear fission products (fall out).

There are promising suggestions that persistent anomalies of the large-scale circulation of the lower atmosphere, affecting the weather of a single season or of longer periods, may be linked with anomalies of the thermal condition of the earth's surface and albedo, which are slow to change in cases where great quantities of heat would be required to eliminate the anomaly: for example, high or low water temperature over great areas (and possibly considerable depths) of the ocean and deficient or excessive ice cover. A proper appreciation of external controls of the atmospheric circulation, whether solar or terrestrial in origin, must be fundamental to understanding the longer-lasting changes in the circulation.

Long series of monthly mean sea-level pressure charts for every January and July since 1750, covering as much of the world as possible, have been produced in the Meteorological Office at Harrow. The methods by which these maps were drawn and limited to regions with a predetermined error tolerance of less than (standard error) ± 2.5 mb. (± 1.0 mb. in summer in the northern hemisphere) have been described elsewhere (Lamb and Johnson, 1959). These maps are being studied for interactions within the circulation itself (for example, between northern and southern hemispheres and between the Atlantic and Pacific sectors of either hemisphere), as well as for relationships between the atmospheric circulation and sea temperature and ice distributions, volcanic dust veils, and any identifiable solar-period effects. The chart material will make it possible to test various current theories about types of circulation change and responses to external events. Applications to the interpretation of recent climatic history and to forecasting the weather of the season ahead (and, in so far as possible recognizing longer-term climatic trends) are obvious ends in view. The possible effects of "random" climatic disturbances in nature (volcanic or other geographical upheavals) or of man-made attempts to manipulate the climatic environment on any scale may also be better envisaged.

Responses of the Atmospheric Circulation to Changes in the Heating Pattern

Lessons of the normal seasonal changes. Circulation developments in response to changing heat gain and loss at the earth's surface and aloft may be studied in relation to the normal seasonal changes as well as in relation to the differences between the same season in different years or epochs. The changes of effective heating between summer and winter greatly exceed the differences between corresponding seasons in different epochs except over periods such that a great ex-

tension of the ice sheets is involved. In so far as we can recognize similar changes of circulation pattern accompanying similar seasonal and secular changes in the heating pattern, we may reach a fundamental understanding of the physical and dynamic responses of the atmosphere to changes in the heat budget. Here both seasonal- and climatic-change studies serve the same meteorological end, and the results may be applied to both problems.

The normal seasonal progression of the circulation over much of the northern hemisphere has therefore been studied in some detail with the aid of successions of pentad normal (5-day mean) maps averaged over 20 years, and 2 different 20-year epochs (1890 to 1909 and 1919 to 1938) have been compared. The lessons may be summarized as follows.

(1) The general atmospheric circulation intensifies as the prevailing latitudinal thermal gradient strengthens to its winter maximum.

(2) The strongest thermal gradient and the associated main depression track move equatorward in winter and poleward in summer, and tend to keep fairly close to the limit of snow/ice-covered surface over those broad sectors where this limit is most stable and least contorted. (This phenomenon appears more clearly and in simpler form in the southern hemisphere for geographical reasons whereas, in the northwestern sector of Eurasia, depressions from the Atlantic commonly pass over the open water of the Barents Sea north of the snow-covered continent before plunging southeast to lower latitudes or filling up over the polar ice.)

(3) Waves in the upper westerlies (reflected in the depression tracks) correspond both to regions of established warm or cold surface and to dynamic effects in the upper westerlies (Rossby waves).

(4) The weakest and most chaotic circulation over much of the northern hemisphere occurs in spring when the thermal distribution in different parts of the temperate zone is most irregular, owing to remaining tongues of snow, ice, and waterlogged surfaces and of cold sea water. Late winter-spring (March to May) at the present epoch is also the season when "blocking anticyclones" in temperate and subarctic latitudes are most frequent, and meridional flow, favored by large amplitude (north-south) waves in the isotherms around the hemisphere, is most marked. There is a second seasonal peak of blocking frequency in the autumn-early winter (October to December) when the difference of temperature between the warm waters of the northern Atlantic and the land surfaces in the same latitudes is greatest.

(5) Weak circulations also arise at the height of summer in those parts of the hemisphere where the heating distribution becomes most uniform.

(6) There is an evident tendency for the circumpolar whirl to shift its center toward the Atlantic-European sector at the height of summer, perhaps because well-heated surfaces on the opposite side of the hemisphere shift the principal thermal gradient (and strongest upper westerlies) in that region to near 70° N; on the other hand, cold water surfaces remain much farther south between 90 and 20° W (in summer the northern hemisphere circumpolar vortex is not so much distorted by large amplitude waves as in winter, and this displacement of the whole circulation toward one sector can readily be discerned).

(7) Upon the thermal effects suggested above are superimposed the dynamic effects of the long waves in the upper westerlies (Rossby waves). Prevailing

wave lengths lengthen as the circulation intensifies, and shorten as the circulation weakens. These waves determine preferred positions for secondary warm ridges and cold troughs around the hemisphere downstream from any geographically fixed ridge or trough.

(8) The most stable (long-lasting) circulation patterns appear to arise when the Rossby wave trains generated downstream from the chief geographically anchored warm ridges and cold troughs happen to "fit" with one another. These stable (standing-wave) circulation patterns and the associated long spells of set weather character break down when the seasonal change of thermal gradient and circulation intensity changes the preferred wave lengths too much for the fit to continue.

(9) At certain times in this seasonal course—for instance when the circulation is strengthening in autumn and early winter—the wave trains set up downstream from the cold troughs established over the cold surfaces of the northern and northeastern parts of Asia and North America apparently begin to interfere with one another. Vigorous depressions tend to transport warm air from the Pacific over Canada or from the Atlantic into the heart of Eurasia. In extreme cases this may delay, displace, or partially suppress the development of the other main cold region with corresponding effects upon the upper cold trough normally developed there. The frequency and vigor of depressions traveling out over the Atlantic or Pacific, as the case may be, should be (and appear to be) similarly affected.

(10) Several of the most regular synoptic sequences in the course of the year seem to start or finish quite abruptly when the main stream of the upper westerlies shifts to a point slightly past (north or south of) or just against some physical barrier (for example, the Himalayas, the Alaskan Rockies and, at times, even the Greenland and Scandinavian mountains). The development of the pressure systems concerned does not proceed smoothly to a seasonal maximum but rather in successive pulses that may have characteristic periods of the order of 15 to 20 and 30 to 31 days.

Responses of the Atmospheric Circulation to Changes in the Heating Pattern

Lessons of the secular changes observable in January since 1750. If we now examine the circulation changes in January, since 1750, using the long series of monthly mean surface pressure charts, we become immediately aware of the nature of some of the climatic changes that have taken place since that time. These are illustrated by FIGURES 4*a* and *b*, in which the average January pressure pattern in an early 40-year period (1790 to 1829) is compared with that used as the modern normal (1900 to 1939), and by FIGURE 5, which reproduces the average January patterns for each decade from the 1760s to the 1950s.

The most obvious differences in the January circulation between the period around 1800 and the first 40 years of the present century are: (1) increased vigor in the present century, with a deeper Iceland low, more intense Azores high, and stronger pressure gradients; (2) a more zonal pattern in the later period, the pressure field for prevailing westerlies in the middle latitudes being a more pronounced feature of the map for 1900 to 1939; and (3) a more southerly position of several of the principal features of the circulation in the Atlantic sector in 1790 to 1829 than in the present century; this applies particularly to

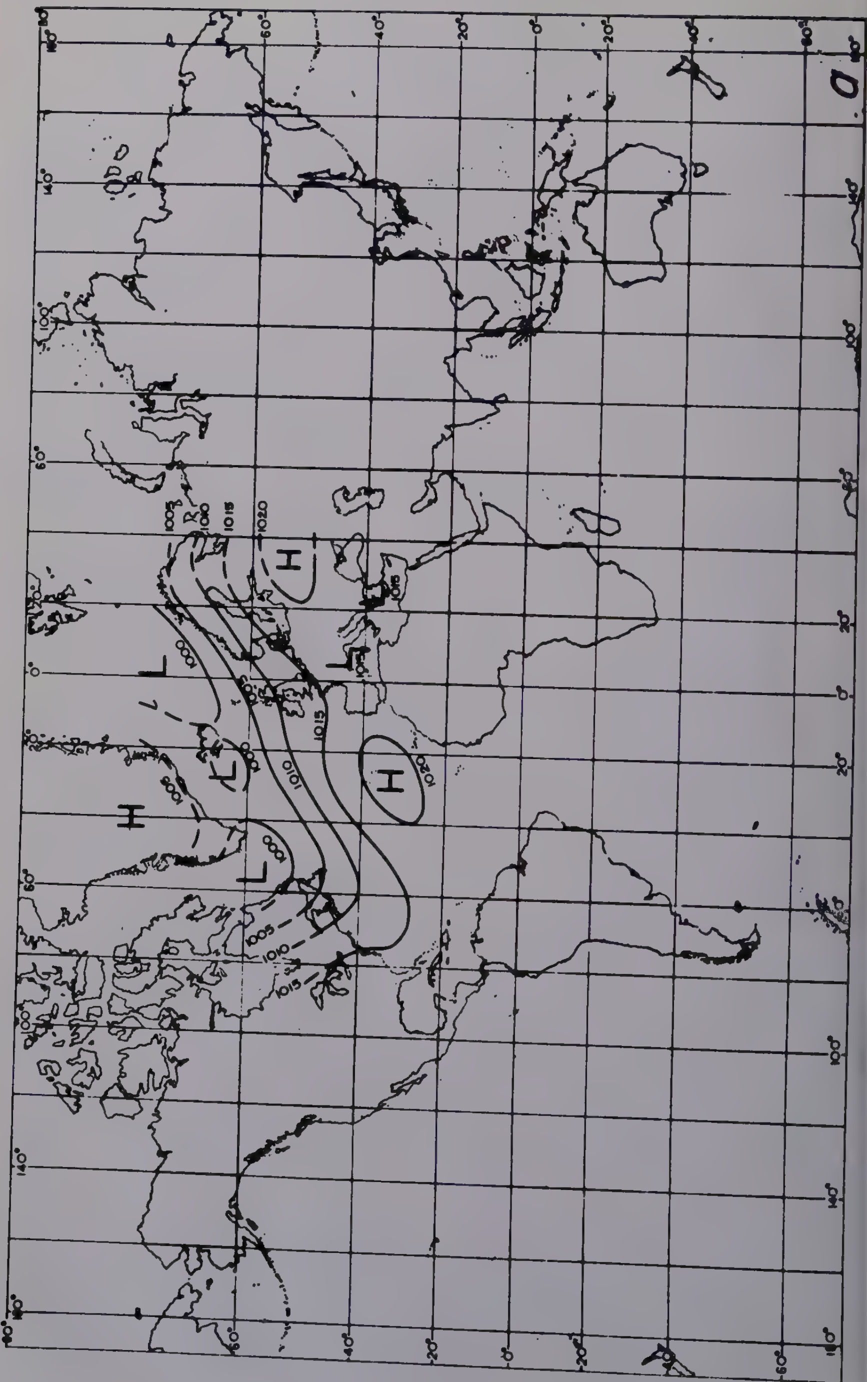
the depression track over the western Atlantic and the large trough extending south off the Atlantic seaboard of North America.

The earlier period was one of generally colder winters and much more extensive sea ice in the higher latitudes. There can be little doubt that the more southerly portion of the depression track where it emerged from North America in the Januarys of 1790 to 1829 corresponded to a more southerly position of the snow and ice limit and of the zone of strongest south-north thermal gradient.

FIGURE 6 displays the differences of ocean-surface temperature between the period around 1800 and that about the early part of the present century, in so far as these changes can be determined from the first sea-temperature survey done for the British Admiralty; this comparison tends to confirm the more southerly position of the main thermal gradient in the earlier period. The features of this sea-temperature-change chart can very largely be explained as a slight southward shift of the main warm-water areas in both the North and South Atlantic in the period around 1800, as compared with their more recent positions. Since the key features of the atmospheric circulation, at least in the North Atlantic, were also farther south than they have been more recently, it is reasonable to suppose that the ocean-current pattern was correspondingly a little farther south around 1800. This would mean that less of the warm equatorial current from the South Atlantic formerly made its way to the Caribbean and ultimately into the Gulf Stream (this is perhaps the explanation of the cooler patch off the coast of the Guianas). The change of sea temperatures in July is remarkably similar to that illustrated for January.

Descriptive evidence from the accounts of some of the navigators and scientists of the time seems to support this postulated southward anomaly of the water-current pattern around 1800, including more southern positions than now of both the arctic and antarctic sea-ice limits; therefore there is some suggestion that the southern hemisphere atmospheric (and ocean) circulation systems were also somewhat farther south, in and near the Atlantic sector, than in recent decades. I have suggested elsewhere (Lamb, 1958) that a slightly lower average world temperature than now would result in more extensive ice on the Arctic seas but less ice in the subantarctic zone because a colder antarctic icecap would flow less and therefore produce fewer icebergs and less (cold) melt-water. A difference of only 1 to 2° C. is suggested, and the displacements of the atmospheric circulation features between 1800 and 1940 probably amounted to only 1 to 3 degrees of latitude in the averages taken over 3 or 4 successive decades. The change in the Arctic ice since 1800 is portrayed in FIGURE 7.

The Pacific sector was probably much less affected. A period of milder winters in Japan between about 1700 and 1890 (Arakawa, 1954), when Europe and eastern North America were having colder winters than in the present century may indicate a slight opposite (northward) anomaly (over that particular period) of the circulation features in the Pacific. It is not possible to be certain about this, because a generally weaker circulation should mean a weaker tendency for westerly winds in Japan and less continental cold air in winter, which would also produce the kind of effect observed. However we shall see later (FIGURES 11 and 12, below) that Japanese winters have not always followed an inverse trend to those in Europe.


a
H
L
1000
1005
1010
1015
1020

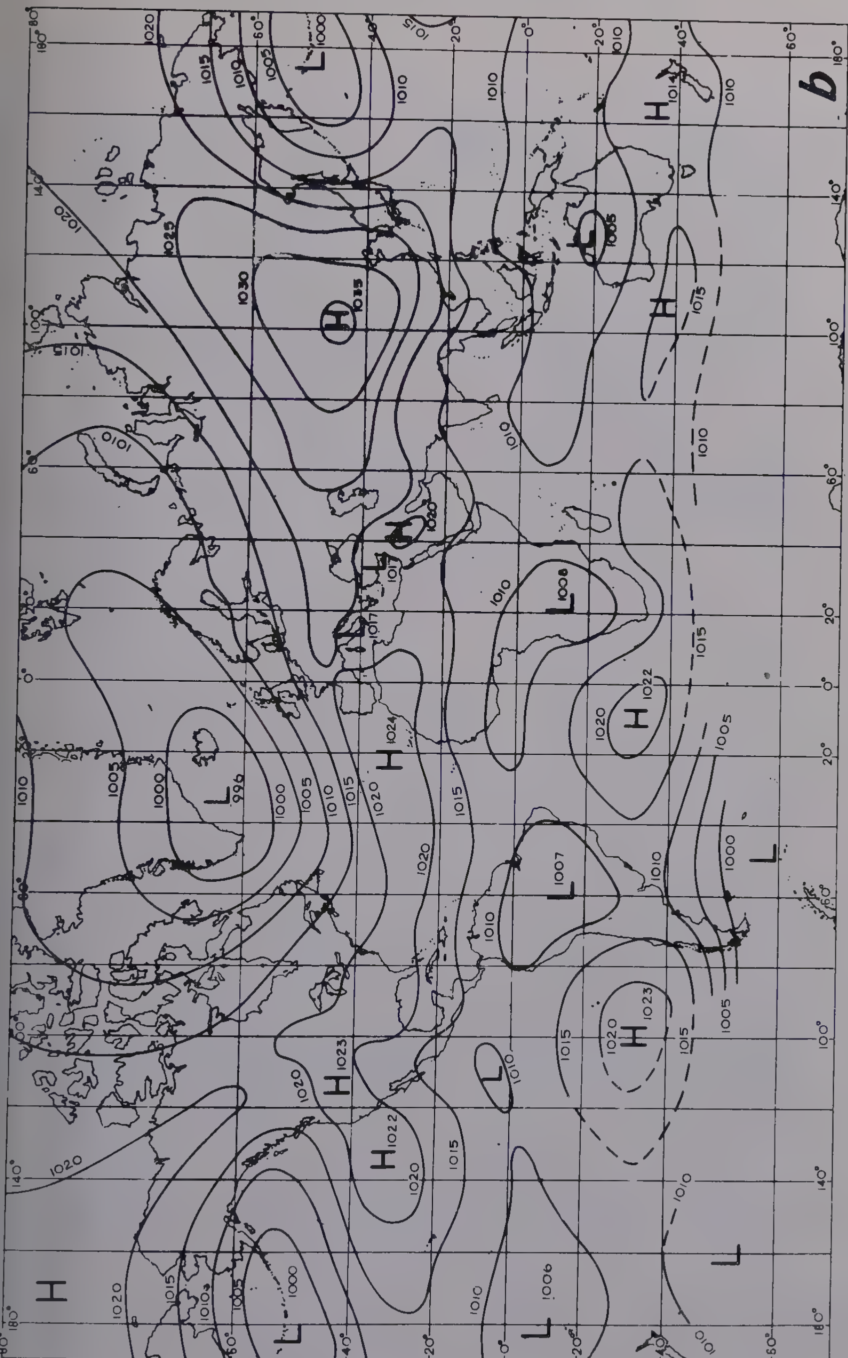
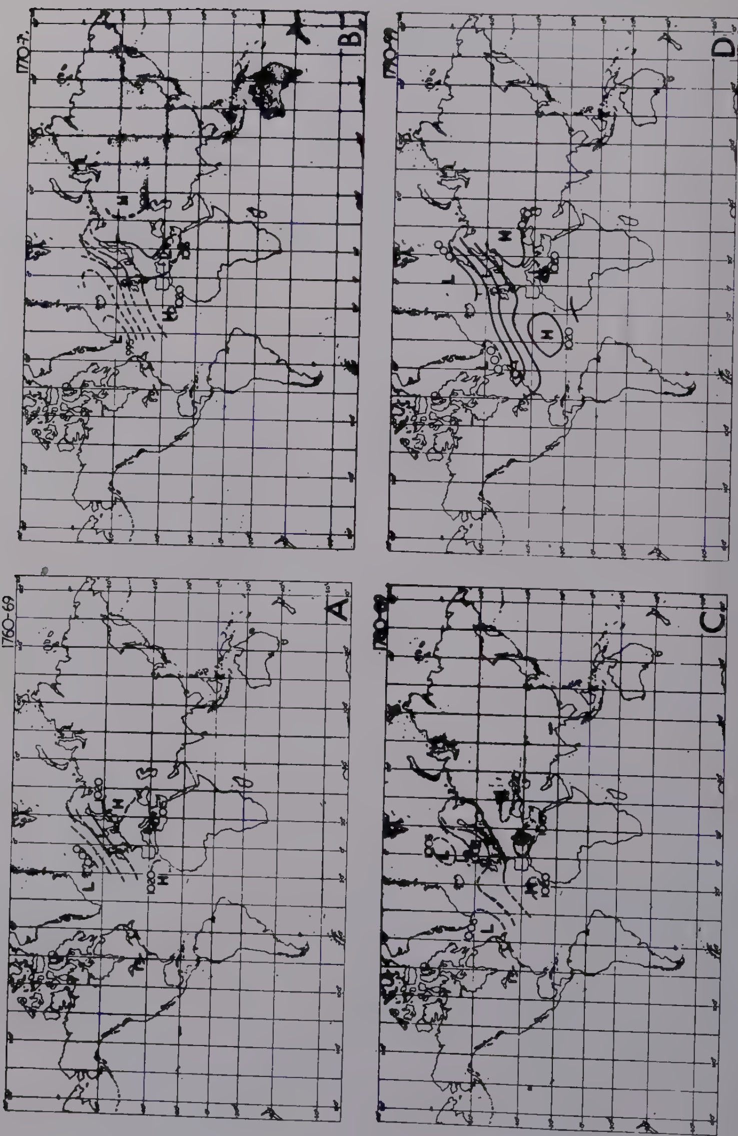
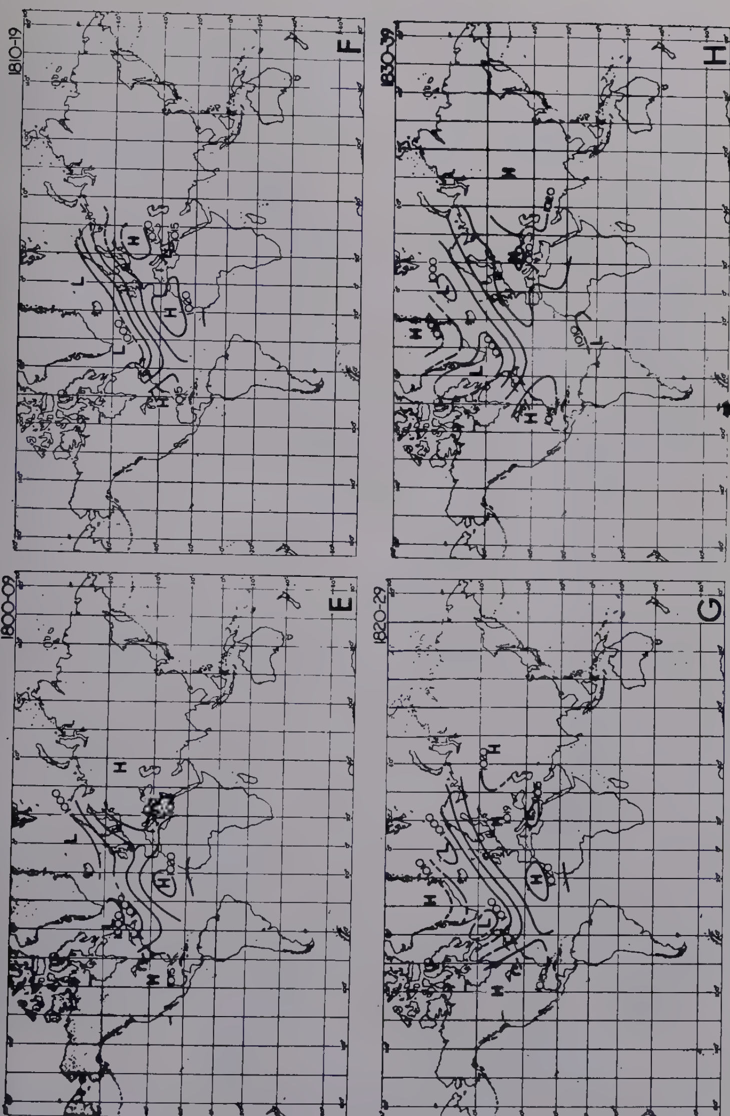


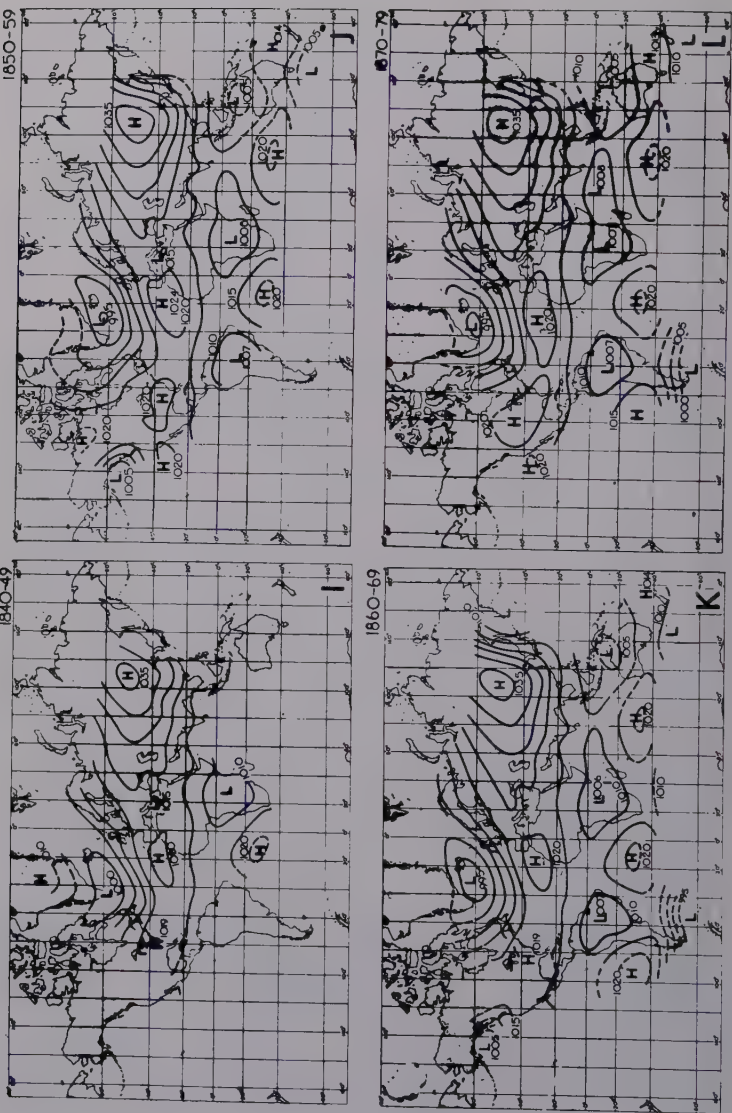
FIGURE 4. The 40-year average mean sea-level pressure distribution for January: Map *a*, 1790 to 1829; map *b*, 1900 to 1939. Reproduced by permission of *Geografiska Annaler*, Stockholm, Sweden.



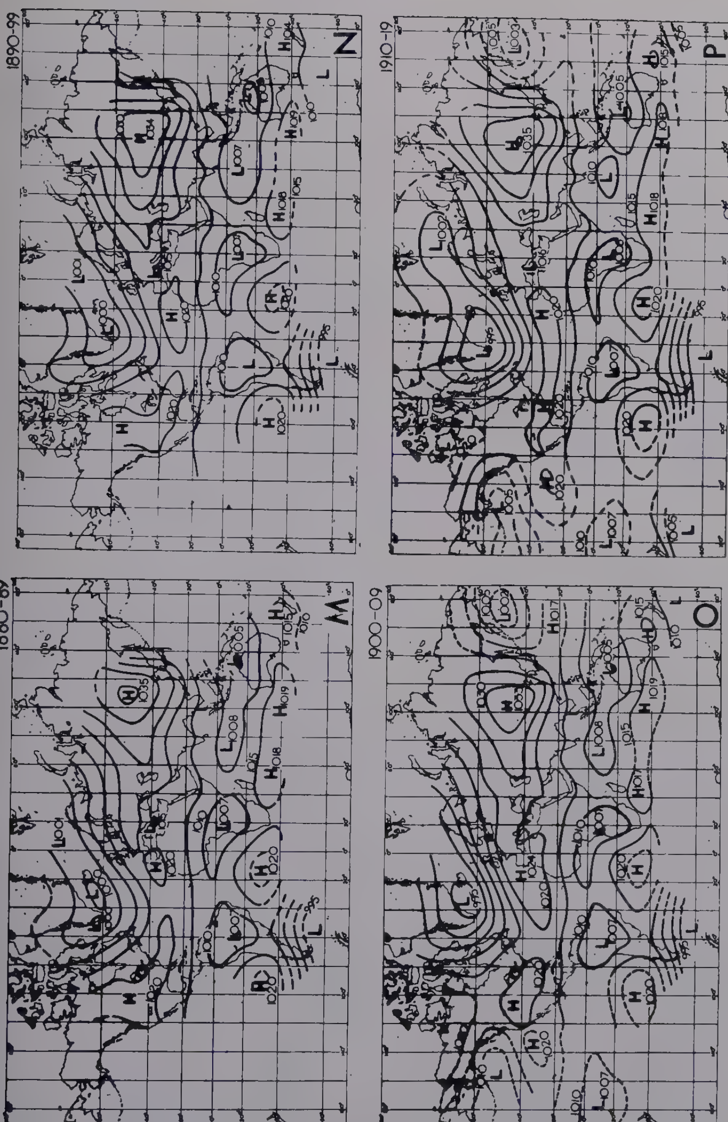
FIGURES 5*a* to *d* (see legend, p. 142).



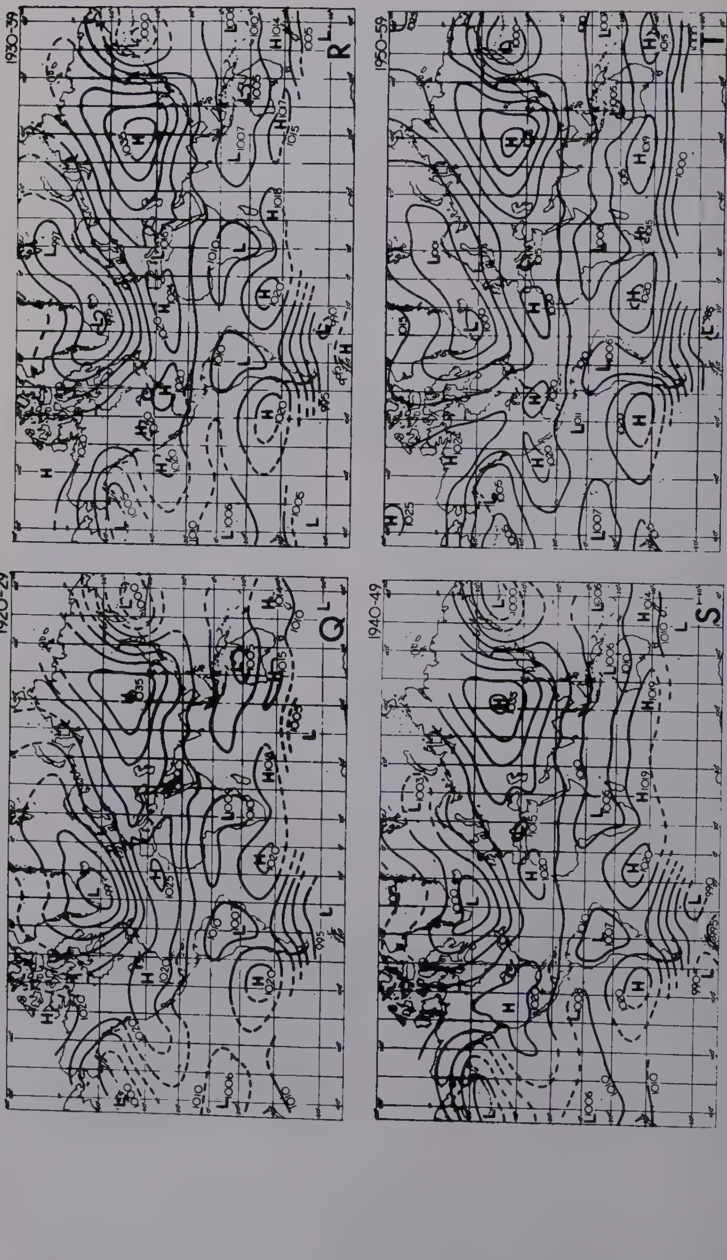
FIGURES 5e to h (see legend, p. 142).



FIGURES 5*i* to *l* (see legend, p 142).



FIGURES 5*m* to *p* (see legend, p. 142).



FIGURES 5*q* to *t*.

FIGURE 5. (*a* to *j*) Average m.s.l. pressure for January, decade by decade, from the 1760s to the 1850s (mb.); (*k* to *t*) the average m.s.l. pressure for January, decade by decade, from the 1860s to the 1950s (mb.). Reproduced by permission of *Geografiska Annaler*, Stockholm, Sweden.

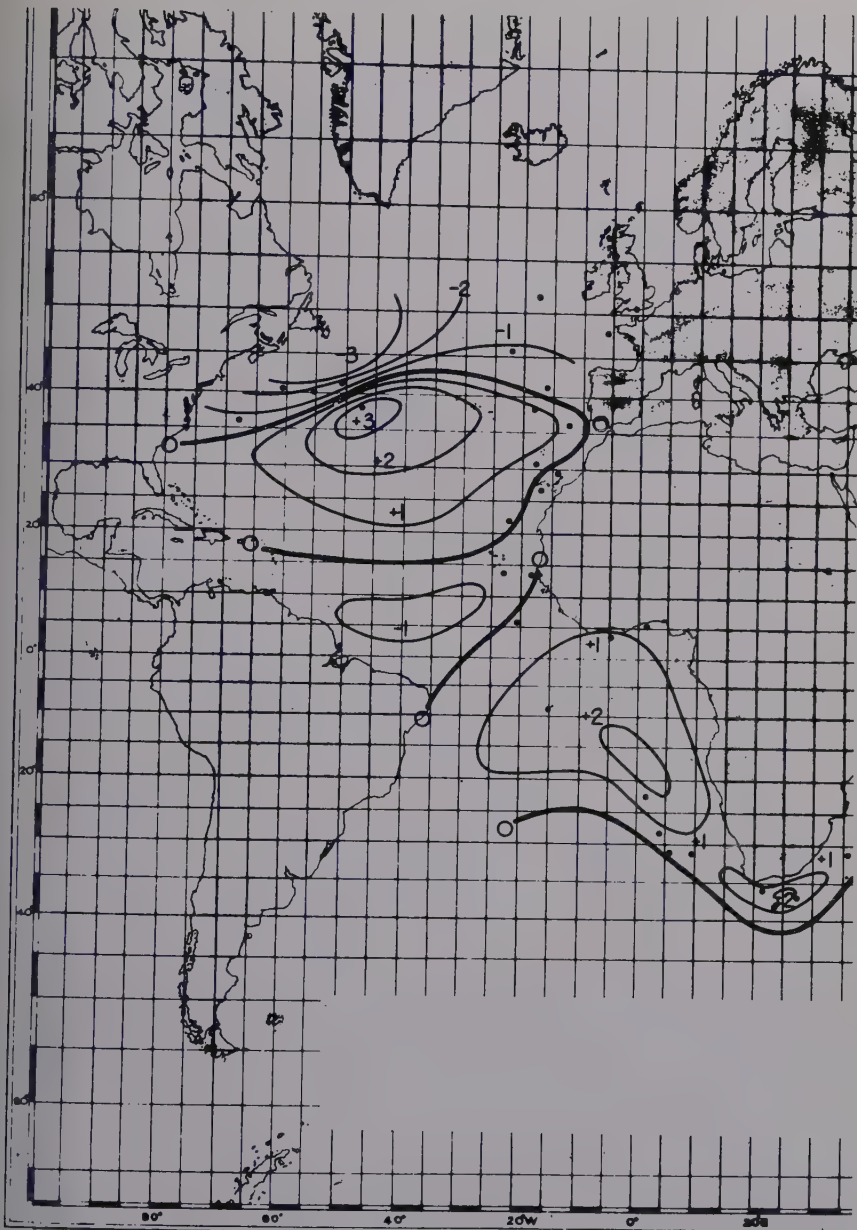


FIGURE 6. Sea temperatures in January 1780 to 1820, expressed as departures from modern values. Sea temperature deviations °C. 1780/1820 from 1887/1899, 1921/1938. Reproduced by permission of *Geografiska Annaler*, Stockholm, Sweden.

FIGURE 8 illustrates the increase of vigor of the atmospheric circulation in January from the early 1800s to the 1920s in the North Atlantic westerlies and trades, and from the 1860s or earlier to the decade 1910 to 1919 also in the southern hemisphere westerlies over New Zealand. There was evidently a world-wide increase in the strength of the zonal circulation of the atmosphere. In both northern and southern hemispheres it appears that the January circula-

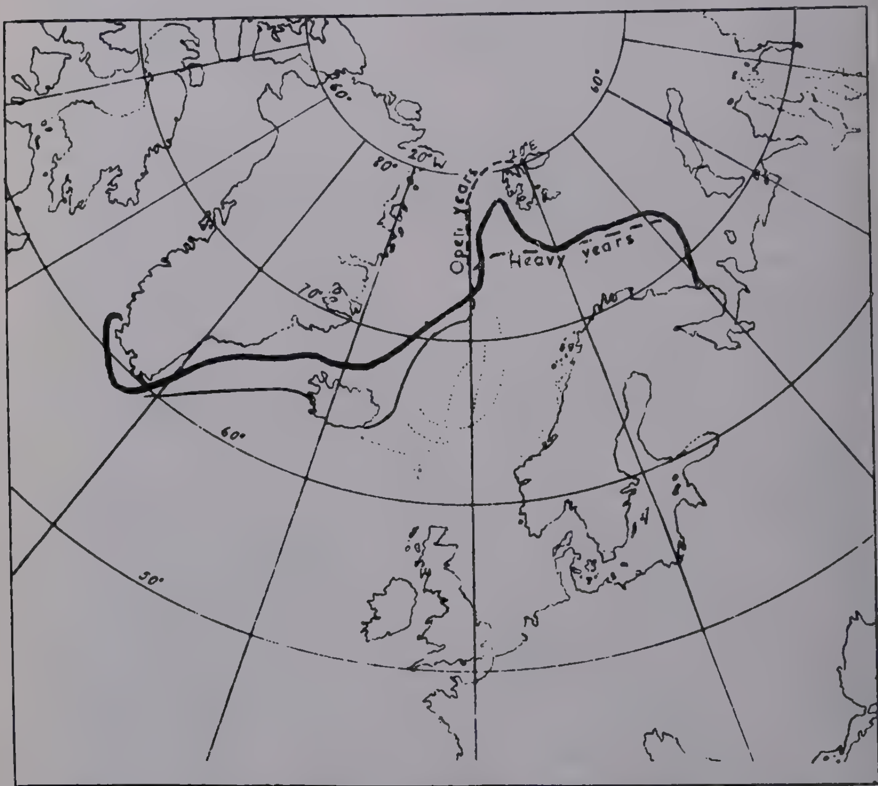


FIGURE 7. Spring maximum extent of arctic pack ice in various years. Key: Heavy line normal (April) 1920 to 1938; thinner line, normal ca. 1800 to 1818 (adapted from W. Scoresby); and dotted line, approximate deduced maximum in extreme years between 1770 and 1820. Reproduced by permission of *Geografiska Annaler*, Stockholm, Sweden.

tion strength has on the whole been declining somewhat over the last 2 to 4 decades.

It may be significant that monthly charts analyzed at Harrow now appear to show a renewed southward trend of the North Atlantic depression track in many months of the year, becoming increasingly prominent in the most recent years (an extreme example is presented in FIGURE 3*a*). Analogues for circulation patterns observed in the last few years are commonly found in the 1880s and 1890s and occasionally in the early years of the last century; seldom between 1900 and 1939. This indicates the relevance of these studies of the

circulation patterns of a century and more ago to the understanding of the present epoch and possible future trends. However, any attempt to foresee climatic trends of the future must be based upon understanding of the physical controls of the atmospheric circulation.

Study of the sequence of decade mean maps of the January circulation (FIGURE 5) reveals fluctuations of strength and position of the main features superimposed upon the rather longer-term trends discussed above; the fluctuations in mean strength of the zonal westerlies since 1820 over the North Atlantic amount to more than 25 per cent between the extreme decades. An interesting chronology could be compiled linking the prominent fluctuations of the circulation with the reported incidence of mild and severe winters and precipitation changes in different parts of the world. Whether any of the fluctuations of circulation vigor are truly periodic oscillations is for further investigation, but there are suggestions—perhaps in the northern hemisphere circulation only—of periodicities of about 20 and 55 to 70 years. Furthermore, it is thought that there are cases in the decade maps (FIGURE 5) where the North Pacific circulation is strong, while the North Atlantic circulation is weak, or vice versa.

The strength of the mean meridional currents in the general circulation in January (FIGURE 8, lower 2 curves, right hand side) does not appear to show any obvious trend comparable with the increased strength of the zonal circulation from about 1800 to 1930. The meridional circulation does however show a suggestion of an approximately 60-year oscillation, which presumably affects the frequency of blocking anticyclones. The curve of mean January pressures at Trondheim, Norway (63° N, 11° E) in FIGURE 9 is suggestive in this connection, but the peaks of this curve appear somewhat out of phase with the peaks of North Sea southerlies. FIGURE 12 also appears to convey suggestions of a similar nature.

To probe for physical controls, it is planned to examine the individual January circulation patterns in relation to (1) similar ocean temperature and sea-ice patterns; (2) similar phases of the 11-year and longer sunspot cycles; and (3) situations in years following volcanic eruptions known to have spread great and persistent dust veils.

The preliminary examination of the January (and July) chart series thus far suggests that factors 1 and 3, above, are important. It is too early to express any opinion on factor 2, apart from impressions gained from other research elsewhere.

The sustained change of vigor of the circulation noted (FIGURE 8) since 1800 amounts to several per cent (between 5 and 10 per cent for 40-year means over the North Atlantic) and evidently demands explanation in terms of possible changes of the available energy supply and possible damping effects arising from the independent circulation of the upper stratosphere and of tropopause height (Kraus, 1960). In this connection an appeal to the existence of more frequent and intense volcanic dust veils in the stratosphere before 1850 or even 1920 than since then is attractive but cannot explain the change fully, for the renewed slight weakening of the atmospheric circulation since about 1930 cannot readily be explained in the same way.

Study of the longitudinal positions and spacing of troughs and ridges in the

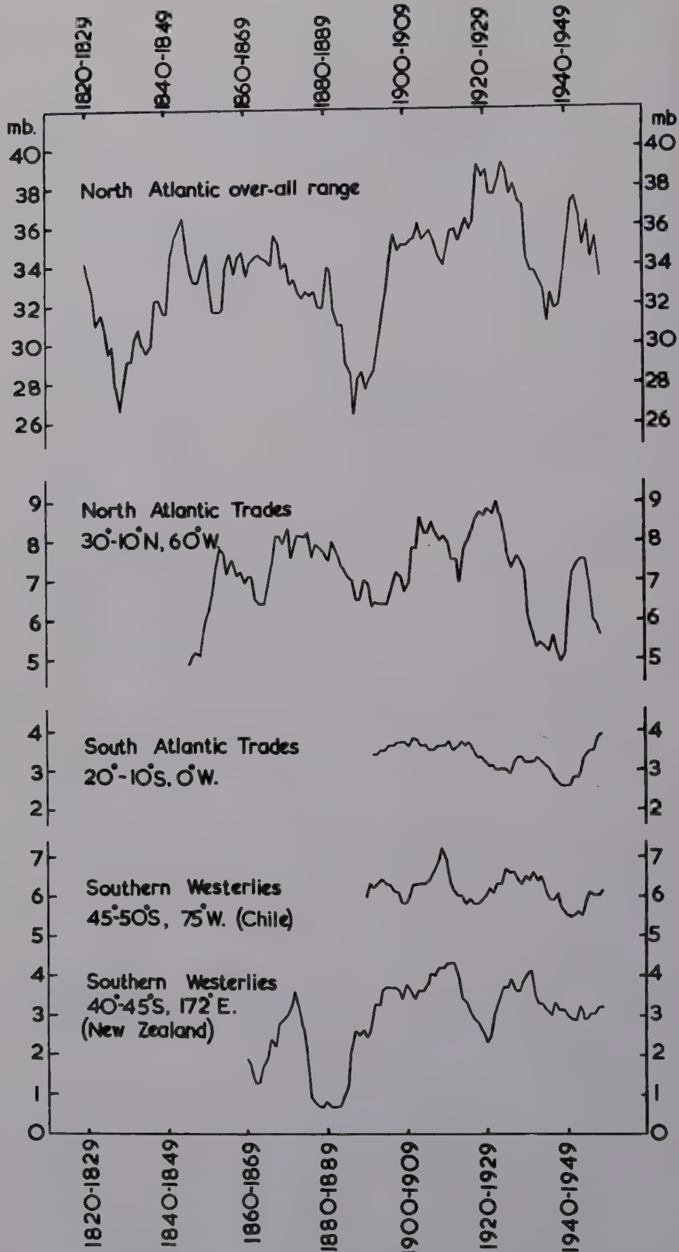
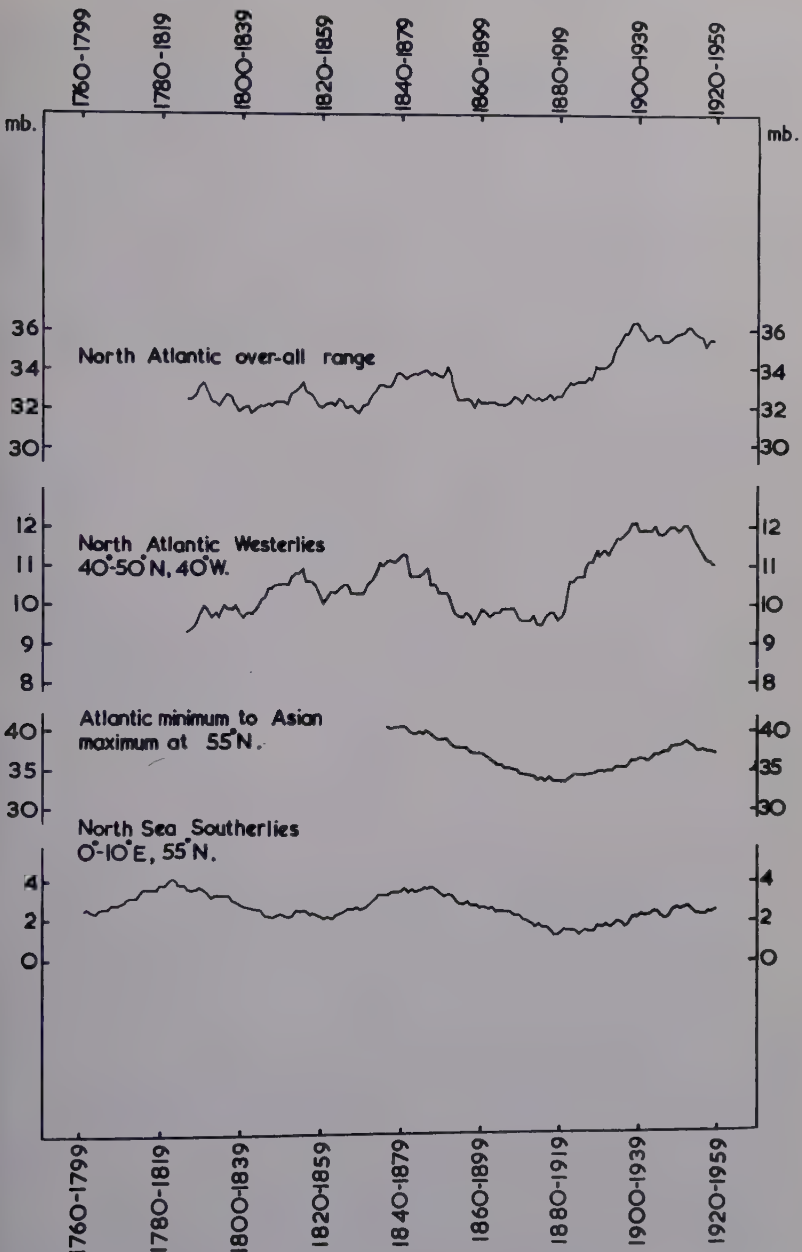


FIGURE 8. Pressure-difference indices of circulation intensity in January. Left graph, *Geografiska Annaler*, Stockholm, Sweden.



10-year running means; right graph, 40-year running means. Reproduced by permission of

belt of surface westerlies, as indicators of the major cold troughs and warm ridges in the upper westerlies, confirm the change in prevailing wave length since 1800 to be expected from the Rossby formula linking stationary wave length (L) with prevailing speed (U) of the zonal westerly winds:

$$U = \beta L^2/4\pi^2$$

where β is the rate of change of the Coriolis parameter with latitude. FIGURE 10 shows the changes in the mean longitude of the surface pressure troughs and

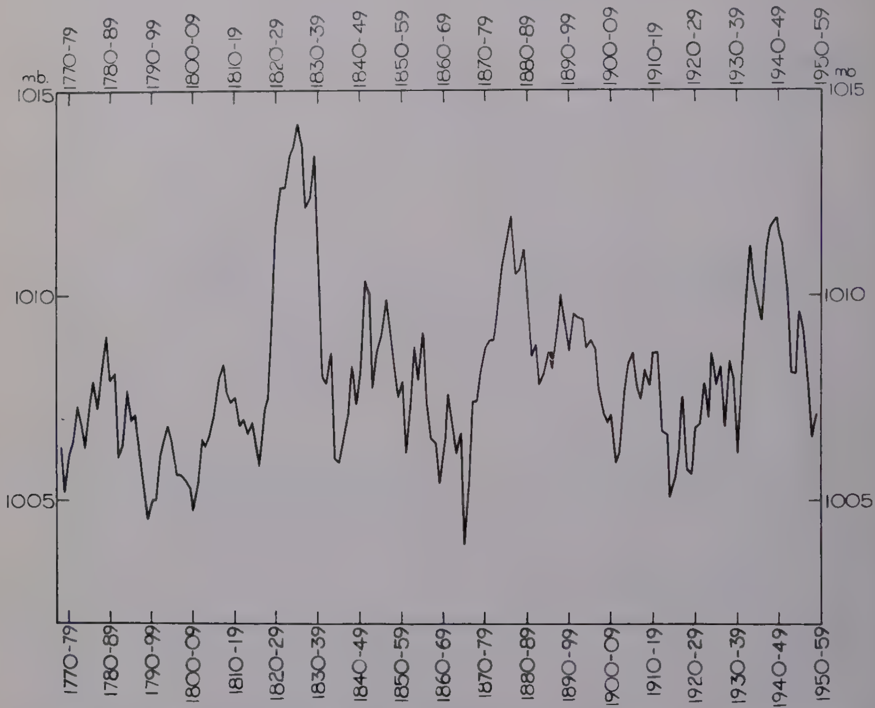


FIGURE 9. Pressure at Trondheim, Norway (63° N, 11° E) in January. Ten-year running means. Reproduced by permission of *Geografiska Annaler*, Stockholm, Sweden.

ridges in January at 45° N (the most reliable part of the older North Atlantic maps). The surface trough over the western Atlantic and the ridge east of it are the features of the surface map at this latitude most obviously linked with the trough farther west in the upper westerlies. Note how these features move east and how their spacing increases from the time of weakest circulation about 1790 to 1829. These trends appear to reverse after about 1900 to 1939. These comparatively minor changes of preferred positions for troughs and ridges can be important for the incidence of cold and warm air streams in the longitudes affected, and may enter into the relative frequencies of cold and mild winters in Japan, western or eastern North America, western Europe, and Russia in different epochs.

In the Atlantic sector climatic changes involving shifts of latitude tend to be

amplified in comparison with other sectors of both hemispheres, presumably because the North Atlantic offers the only effective outlet for increased supplies of Arctic sea ice and the only major access for warm water to the Arctic. A contributory factor may be sensitivity of the surface-water temperature distribution to the proportion of the warm equatorial current dividing at the "nose" of Brazil near 5° S.

There is clearly much interest in attempting to determine whether any particular longitudes tend to be affected first or more than others by various climatic changes. This may of course depend upon whether the change is induced directly by a change in the available radiation supply (and correspond-

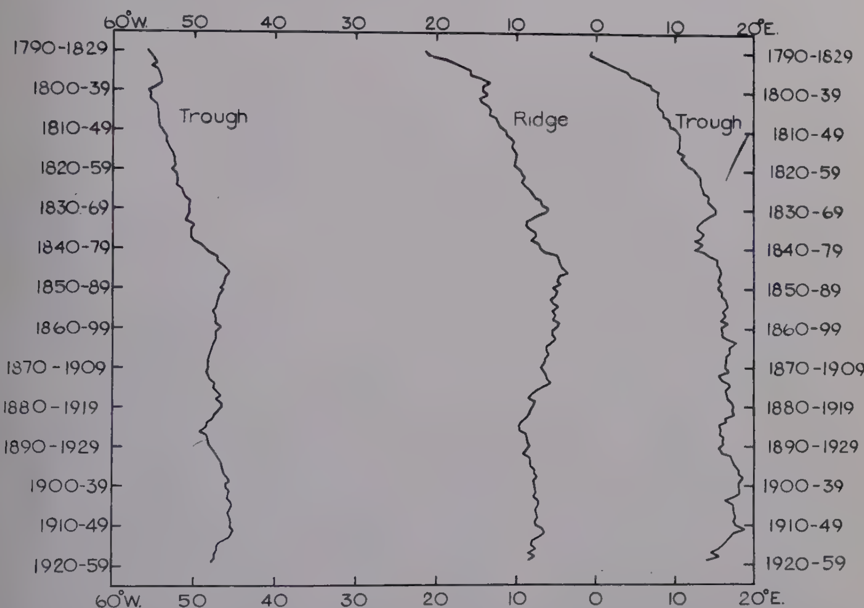


FIGURE 10. Longitudes of the semipermanent surface pressure troughs and ridges at 45° N in the Atlantic sector. Forty-year running means. Reproduced by permission of *Geografiska Annaler*, Stockholm, Sweden.

ing change of general temperature level in the lower troposphere), or whether the immediate cause is some change in the heat stored in the oceans and supplied by the North Atlantic or North Pacific drift currents to the higher latitudes.

Preliminary Analysis of Documentary Weather Data for Winters During the Historical Period Since 1100 A.D.

It should be instructive to survey the documentary records of the historical period before the time of meteorological instruments with a view to discovering how far all longitudes seem to have experienced parallel changes of fortune toward warmer or colder epochs. It may be worth remembering that little more than a generation ago it was commonly denied that the available records indicated anything more than a haphazard incidence of good or bad years and

natural disasters and that, outside of Scandinavia, few historians seem to have considered the frequencies of different kinds of reported weather phenomena.

The following compilations of documentary records of weather character in different months or seasons cover the centuries since 1100 (records of weather before that date are much sparser and appear inadequate for our present purpose).

Iceland, 1291 to 1392 only	(Bull, 1916)
France	(Angot, 1895)
Belgium	(Vanderlinden, 1924)
Western Europe, generally	(Easton, 1928)
Europe generally, especially central Europe and Italy	(Hennig, 1904)
Baltic, ice conditions only from 1530	(Betin, 1957)
Russian plain, especially Ukraine	(Buchinskiy, 1957)
Japan, freezing dates of Lake Suwa	(Arakawa, 1954)

Crude numerical indices of winter severity over various periods of years may be derived from these records by taking (1) the frequency ratio of indications of severe or obviously cold to obviously mild winter months (December, January, and February); and (2) the excess number of unmistakable mild or cold winter months recorded. The first of these, index A, is generally regarded as more reliable than the second type, index B, because the effects of changes in the amount of attention devoted to weather in the annals in different periods are eliminated. Any supposed bias on the part of the annals in the direction of taking more notice of severe than of warm winter months should be of little importance when values of the same index (whether A or B) in different periods are compared. On the other hand, the ratio index A is unstable and tends to give large values if the recorded mild winters are too few in number: this index cannot be used at all for comparing periods of less than half a century without the value occasionally running away to infinity in Russia.

FIGURE 11 shows the result of plotting 50-year values of the winter severity ratio index A for Great Britain, western and central Europe, and the Russian plain, all near 50° N, from the year 1100 to 1800, tentatively continued to 1950 with the aid of temperature records and taking account of the continuously comparable Baltic ice record (dates of opening of the port of Riga, Latvia) that, together with a "temperature coefficient" devised by Easton (1928), overlaps the era of descriptive documentary records and the era of thermometers (and is the most satisfactory bridge between them so far available). Additionally half-century average values of the freezing date of the small, rather high-level Lake Suwa in central Japan are plotted in the form of number of days, early (negative) or late (positive numbers), against the over-all mean (January 15): schematic isopleths have been drawn to suggest respectively cold anomalies advancing from the west (Asia) and warm anomalies advancing from the east (Pacific Ocean), since these anomalies probably imply respectively an increase or decrease in the advection of cold air from the continent of Asia.

The compounded frequency of anomalous months, both warm and cold, indicated in the records for various European longitudes was also examined. In western and central Europe the numbers were fairly consistently about 50 mo. per half century—that is, one third of all the winter months—suggesting

that the standard deviation of temperature is approximately the limit beyond which a month secured mention as cold or warm. Any tendency for preferential attention to notably cold rather than notably mild winter months in the annals appears small enough to be neglected in comparisons between dif-

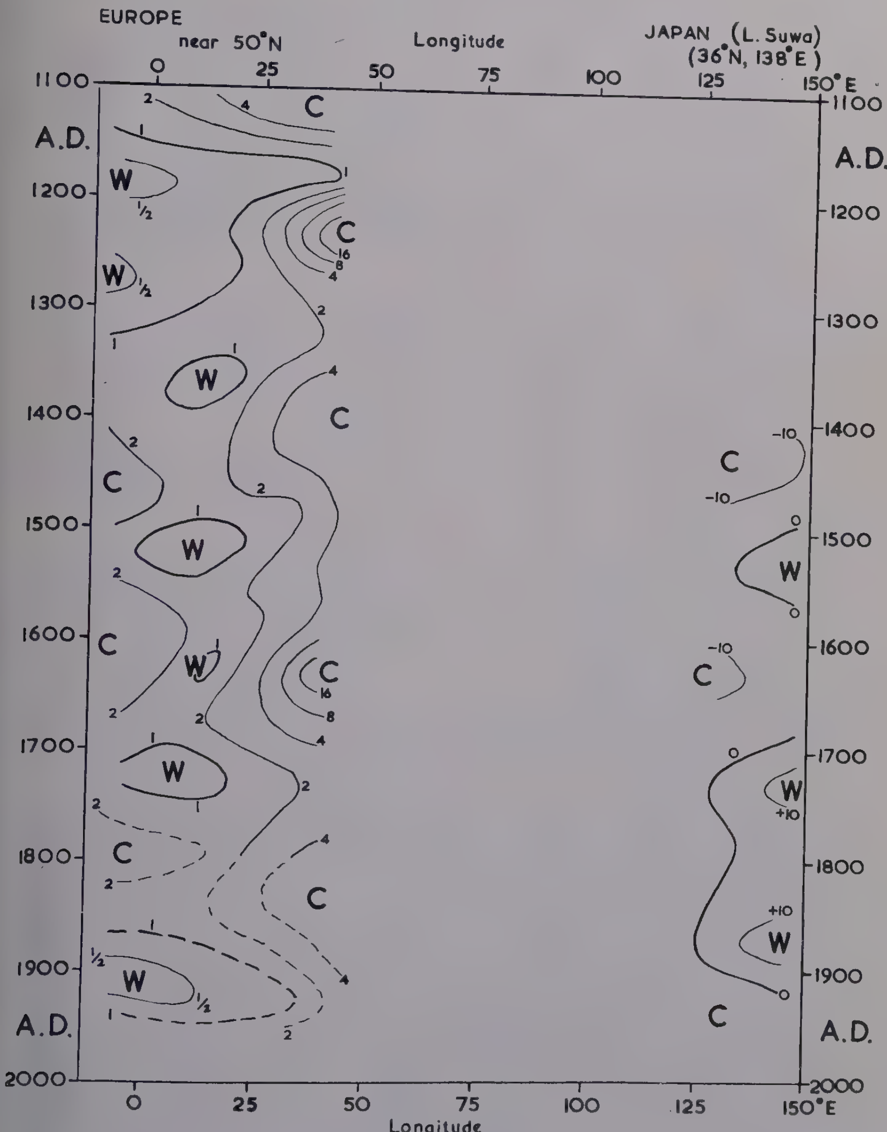


FIGURE 11. Survey of historical records by half centuries (schematic): 50-year values of winter severity (ratio) index A in different longitudes in Europe (frequency ratio of clearly cold to clearly mild months recorded, December, January, and February only), also freezing dates of Lake Suwa, Japan, as departures from the over-all means. Key: — sign = days early; + sign = days late. Reproduced by permission of *Geografiska Annaler*, Stockholm, Sweden.

ferent periods. In Russia at most times only about one-half as many months were recorded as notably warm or cold, and the reported effects upon human life and communications indicate correspondingly greater extremes in the general run of cases mentioned. The values of the ratio (index A) and the number of months indicated as warm or cold are given in TABLE 1. The greater numbers of anomalous or extreme months after about 1500 cannot be wholly attributed to increase of recording: it seems probable from the nature of the records that the variance of temperature also increased in the centuries that followed. We shall see later from index B that a material change in the atmospheric circulation may be deduced about the same time.

The indications in TABLE 1 of an approach to a constant standard of reporting of noteworthy months, warm or cold, down the centuries since 1100, at least in western and central Europe, suggest that it may be legitimate to use

TABLE 1

WINTER SEVERITY INDEX A: FREQUENCY RATIO OF COLD/MILD WINTER MONTHS

Period	SE Britain	Central Europe	Russian plain
1100 to 1149	1.73 (41)*	2.64 (51)	6.0 (7)
1150 to 1199	0.53 (29)	0.62 (39)	0.65 (23)
1200 to 1249	0.79 (61)	0.95 (50)	12.0 (13)
1250 to 1299	0.60 (48)	0.94 (69)	3.0 (12)
1300 to 1349	1.16 (54)	1.20 (73)	1.0 (6)
1350 to 1399	1.08 (50)	0.93 (54)	6.3 (22)
1400 to 1449	1.67 (55)	1.67 (59)	5.2 (56)
1450 to 1499	2.38 (44)	1.80 (69)	2.0 (33)
1500 to 1549	0.92 (50)	0.71 (94)	3.0 (12)
1550 to 1599	2.34 (60)	1.74 (55)	3.0 (20)
1600 to 1649	2.52 (81)	0.98 (48)	16.5 (35)
1650 to 1699	1.61 (107)	2.12 (80)	5.9 (48)
1700 to 1749	0.80 (95)	0.89 (58)	2.0 (27)

* The figures in parentheses are the total numbers of months indicated in the records as anomalous, whether warm or cold.

index B (excess number of notably mild or cold months) to study the character of the successive decades. This is done in FIGURE 12, in which an excess of notably mild months is shown by hatching, for all available longitudes near 50° N. Easton's "temperature coefficient" and the opening dates of the port of Riga have been used, as in FIGURE 11, for amplification of the evidence (although it should be remembered that ice conditions at Riga probably depend partly on the severity or otherwise of the winter in another latitude, that of the northern Baltic in 60 to 65° N). The extension of the index to 1960 remains somewhat tentative until comparisons with instrumentally measured temperatures are more firmly established; nevertheless the outlines of the picture presented in FIGURE 12 can be broadly accepted. The average freezing dates of Lake Suwa, Japan, in each decade are also shown.

TABLE 2 gives the detailed figures used decade by decade from 1100. The index B figures for Russia have been doubled to equalize approximately the rate of reporting of anomalous months (warm or cold) with the figures for western and central Europe. Of the figures contained in the table, Easton's

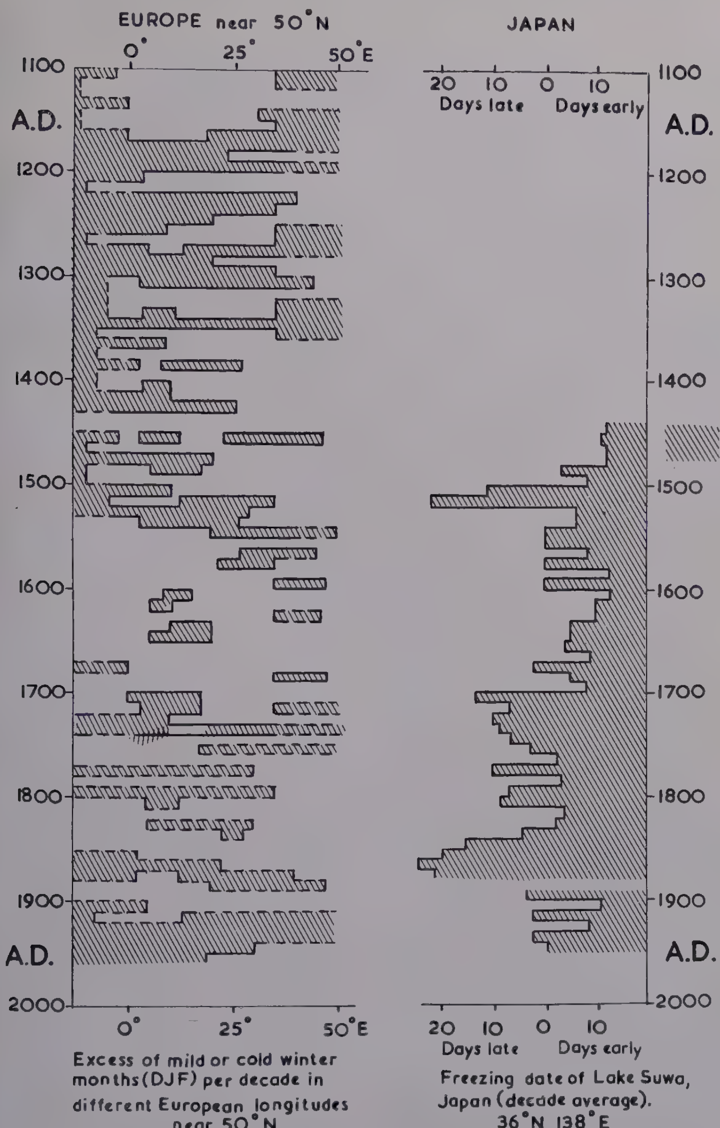


FIGURE 12. Survey of historical records by decades: winter severity index B (excess number of clearly mild over clearly cold months recorded, December, January, and February only) in different longitudes in Europe, also anomalies of the freezing date of Lake Suwa, Japan. A broken outline is used to indicate a possible (slight) change of zero after temperature measurements came into use and other points of uncertainty. Left, excess of mild or cold winter months (DJF) per decade in different European longitudes near 50° N. Right, freezing date of Lake Suwa, Japan (decade average) 36° N, 138° E. Hatched areas indicate excess of mild months. Reproduced by permission of *Geografiska Annaler*, Stockholm, Sweden.

Table 2

Winter Severity Index B and Some Comparative Figures

Decade	Index B + mild − cold			Easton's "temperature coefficient" (western Europe ca. 5° E) 50+	Mean date of opening of the port of Riga (departures from mean date April 5, N.S.) *sign inverted* i.e. + early − late	Mean date of freezing of Lake Suwa, Japan day 15+ (i.e. departures from mean date January 15) − early + late
	Britain ca. 0° E	Germany ca. 12° E	Russia ca. 35° E			
1100 to 1109	−1	−2	0			
1110 to 1119	−4	−5	0			
1120 to 1129	−3	−3	−6			
1130 to 1139	0	−3	−6			
1140 to 1149	−3	−9	+2			
1150 to 1159	−6	−3	0			
1160 to 1169	0	−2	+6			
1170 to 1179	+2	+2	+4			
1180 to 1189	+6	+5	−6			
1190 to 1199	+7	+5	+6			
1200 to 1209	+1	−3	−6			
1210 to 1219	−2	−5	−12	−13.8		
1220 to 1229	+1	+6	+2	+3.4		
1230 to 1239	+6	+1	0	−4.5		
1240 to 1249	+1	+3	−6	+4.0		
1250 to 1259	+3	−5	0	−3.3		
1260 to 1269	−3	−1	0	+6.5		
1270 to 1279	+7	0	+6	−4.3		
1280 to 1289	+3	+3	−18	+5.0		
1290 to 1299	+2	+5	0	+0.8		
1300 to 1309	−1	+1	+6	+2.0		
1310 to 1319	−1	−10	−6	−4.9		
1320 to 1329	0	−1	0	+0.6		
1330 to 1339	−4	−1	0	+3.7		
1340 to 1349	+2	+4	0	+3.6		
1350 to 1359	−3	−4	0	−2.9		
1360 to 1369	+4	−2	−6	+1.5		
1370 to 1379	−3	−5	−12	−0.7		
1380 to 1389	+2	+11	−8	−0.8		
1390 to 1399	−2	+2	−6	−1.4		
1400 to 1409	−3	−2	−22	+1.3		
1410 to 1419	+2	−1	−10	+2.4		
1420 to 1429	+4	+7	−6	+1.1		
1430 to 1439	−12	−18	−10	−5.4		
1440 to 1449	−5	−2	−28	+0.3		−11.8
1450 to 1459	−1	−3	+8	+3.2		−11.2
1460 to 1469	−9	−6	−6	−4.6		−12.2
1470 to 1479	+3	+1	−6	+7.9		−11.7
1480 to 1489	−6	+1	−10	+1.0		−3.2
1490 to 1499	−5	−12	−8	−9.9		−8.1
1500 to 1509	+7	−3	−6	+1.1		+11.0
1510 to 1519	−2	+2	0	−2.0		+22.3
1520 to 1529	+2	+11	−6	+5.9		−6.6
1530 to 1539	−3	+6	−6	+8.4		−5.7
1540 to 1549	−2	−1	+6	−5.1		+0.4
1550 to 1559	−1	−3	−12	+0.1		+0.2
1560 to 1569	−7	−3	+4	−6.5	0	−8.2
1570 to 1579	−3	−1	0	−1.9	+1	−0.4
1580 to 1589	−3	−5	−12	−15	−1	−11.8
1590 to 1599	−10	−5	0	−3.7	−6	−0.4

TABLE 2—*Continued*

Decade	Index B + mild − cold			Easton's "temperature coefficient" (western Europe ca. 5° E) 50+	Mean date of opening of the port of Riga (departures from mean date April 5, N.S.) *sign inverted* i.e. + early − late	Mean date of freezing of Lake Suwa, Japan day 15+ (i.e. departures from mean date January 15) − early + late
	Britain ca. 0° E	Germany ca. 12° E	Russia ca. 35° E			
1600 to 1609	−6	0	−6	−0.6		−11.7
1610 to 1619	−2	−3	−10	+1.6	0	−9.5
1620 to 1629	−9	−3	0	−4.7	−5	−10.2
1630 to 1639	−6	+2	−10	−1.3		−5.3
1640 to 1649	−12	+5	−36	−1.5		−5.3
1650 to 1659	−9	−7	−24	−6.7	Very variable	−4.1
1660 to 1669	−7	−3	−14	−4.7		−8.8
1670 to 1679	0	−6	−24	−8.3		+2.2
1680 to 1689	−2	−7	0	−1.7		−5.1
1690 to 1699	−7	−5	−6	−8.6		−7.8
1700 to 1709	0	+1	−6	+16.1		+13.2
1710 to 1719	−2	+1	0	+4.9	−4	+7.0
1720 to 1729	+8	−1	−12	+1.6	0	+10.3
1730 to 1739	+11	+3	+6	+0.7	+3	+8.6
1740 to 1749	−6	−1	−6	−10.0	−10	+6.7
1750 to 1759		−5	+10	−5.5	+14	+2.8
1760 to 1769			−12	−2.1	−1	−2.5
1770 to 1779			−6	+3.0	+9	+9.8
1780 to 1789			−20	−8.4	−5	−2.7
1790 to 1799			0	+2.1	+2	+6.5
1800 to 1809				+2.3	−6	+8.3
1810 to 1819				−2.7	−7	−3.8
1820 to 1829				+1.5	+2	−2.0
1830 to 1839				−1.5	+1	+4.0
1840 to 1849				−1.2	−4	+14.5
1850 to 1859				−2.2	−5	+19.0
1860 to 1869				+3.7	0	+24.0
1870 to 1879				−1.8	+4	+20.5
1880 to 1889				−4.1	+14	
1890 to 1899				−5.7	+2	+2.6
1900 to 1909				−2.1	+1	−11.4
1910 to 1919					+12	+2.3
1920 to 1929					+11	−8.8
1930 to 1939					+12	+1.7
1940 to 1949					+5	−1.0
1950 to 1959					−3	

temperature coefficient is the least consistent and has been given correspondingly less weight than the other figures in this analysis: this coefficient has been previously criticized by other authors.

FIGURES 11 and 12 reveal that it is by no means a rule that all European longitudes are similarly affected. Correlation coefficients between the winter severity index B in Great Britain and Germany and Great Britain and Russia were only +0.45 and +0.31 respectively, over the period 1100 to 1750 (65

pairs of values); these figures appear statistically significant, but in both cases the correlation coefficient was negative over parts of the period lasting a century or more. In this connection a finding of Betin's (1957) from comparisons of Baltic ice amounts and Caspian Sea levels, which depend largely on the discharge of the River Volga is interesting: in the present century these two items are negatively correlated, and anticyclones that give cold winters in the Baltic have tended to cover the Volga basin also, whereas in certain earlier epochs the opposite is true and anticyclones over Scandinavia were apparently usually accompanied by cyclonic conditions over the Volga basin and vice versa.

There seems to have been rather more tendency for anomalies of the same sign across most of Europe between 1150 and 1300 or 1350 and since about 1850 than at most other times. These have been the periods with most predominance of mild winters (presumably attributable to westerly winds sweeping far into Europe), although a few decades were cold everywhere (suggesting easterly winds and blocking patterns between 0 and 30° E). In other periods and centuries, cold tongues show a preference alternately for one longitude and then another, suggesting a weaker zonal circulation (probably weaker at times—for example, in the 1600s—even than that between 1800 and 1829; FIGURES 4*a*, 5, and 8) and rather more prominence of northerly and southerly components. The periods of coldest winters between 1530 and 1700, and in a few later decades, appear to have been particularly marked in Great Britain and Russia, suggesting an important role played by northerly outbreaks in two main channels: the Norwegian Sea and the Russian plain, especially the former, since some decades in this period had a surplus of mild months either in central Europe or Russia although never in England.

This goes a long way toward explaining the extraordinarily rapid and great increase in the incidence of sea ice in the Iceland sector. Sea temperatures near Great Britain ultimately fell considerably below present normals (FIGURE 6): the virtual extinction of the codfishery from the Faeroe Islands in the 18th century probably indicates that the 2° C. sea-water isotherm (which is critical for cod) commonly lay near the Faeroes, and a tongue of the polar ice is said to have approached on at least one occasion.

It should also be noted that Soviet workers have found that the coldest east-wind situations in winter over the Baltic and central Europe are those supplied by the northerly outbreaks over the Russian plain that originate east of Spitsbergen and over the Kara Sea east of Novaya Zemlya.

Comparisons between the winter severity index B in Great Britain or Russia and the decade mean figures for Japan (freezing dates of Lake Suwa) yielded correlation coefficients of +0.31 and +0.21 respectively over periods from 1440 to 1750 or 1800 (31 and 36 pairs). Neither of these figures is statistically significant above the 10 per cent level and both cases yielded negative correlation coefficients in certain centuries, indicating that the changes of winter character in Europe and Japan are sometimes inverse over quite long periods lasting up to a century or more.

The following chronology suggests itself from examination of FIGURE 12.

From 1100 to ca. 1170. There were small excesses of cold months in most decades, particularly in central Europe, and a rather milder tendency in Russia.

Possibly there was a rather weak zonal circulation and northerly components presumed rather prominent in the North Sea and Baltic, while southerlies were more probably prominent over the eastern Atlantic and Russia.

From 1170 to 1530. There was evidence of strong zonal circulation over the Atlantic and Europe in the earlier part of this period; apparent general weakening of this zonal circulation occurred from about 1300 onward. Mild months predominated in all longitudes, presumably due to westerly winds, particularly in certain decades at about 50-year intervals: namely (1130s), 1170s to 1190s, 1220s to 1240s, 1270s to 1290s or a little later, 1340s, (1380s), (1410 to 1420s), (1470s), and 1500 to about 1530. In some of the intervening decades most or all European longitudes experienced a surplus of cold months, particularly eastern Europe where the severity of the winters was often disastrous: this suggests north European blocking anticyclones prominent in the decades concerned (especially 1210 to 1219, 1310 to 1319, 1400 to 1409, 1430s, and 1490s).

From 1530 to about 1720. Cold months were predominant, especially in western Europe and Russia, although sometimes in all European longitudes as in the 1550s, 1580s, 1600 to 1630, 1670s, and 1690s. Presumably the zonal component of the circulation over Europe was generally weak, except in a number of well-marked easterly winters; northerlies evidently were prominent in the Norwegian Sea and Russian sectors. This analysis lends credence to a strange report quoted by Brooks (1949, p. 312) from a book dated 1634 and attributed by him to the late 1500s and early 1600s: ". . . old shippers of the Netherlands affirming that they have often noted the voyage from Holland to Spaine to be shorter by a day and halfe sayling than the voyage from Spaine to Holland." An abrupt change is noticeable about 1700 to 1720 when predominance of mild months in western Europe suggests frequent southerly components. This was followed by one decade, the 1730s, of notably mild conditions in all European longitudes, probably marking a strong thrust of Atlantic westerlies sweeping across the continent.

About 1750 to 1900. After 1 decade, the 1740s, which was predominantly cold in all European longitudes (easterly winds prevailing?), this was an epoch of generally weak zonal circulation over the Atlantic and temperate Europe (FIGURES 5 and 8), northerly and southeasterly cold air streams each being important at times. Winters were mostly colder than now. There were some decades of stronger zonal circulation, for examples 1750s to 1770s, 1790s, the period 1817 to 1825, and the 1850s to 1870s; and easterly winds were especially prominent in certain years in the 1780s, 1830s, and 1880s to 1890s. The latter half of the 19th century apart from the last 2 decades clearly belongs to the period of strengthening Atlantic westerlies, as is seen from the indices measuring pressure gradients (FIGURE 8).

Since 1900. This was a period of strong Atlantic westerlies, especially in the first decade and the 1920s and 1930s, leading to a decade, the 1940s, when westerlies and easterlies were each pronounced over Europe in different years. Since 1950 the zonal circulation in the Atlantic-European sector appears to have been weaker, with only 1 notably mild winter, although there were several mild (or even warm) Decembers and Februarys in Europe in which southerly components played the main part; the severe months of this last decade mostly

affected a rather narrow range of longitudes in either western, central, or eastern Europe, and were associated with northerly winds in the sector concerned.

Conclusions

The types of analysis of seasonal and secular changes of weather, climate, and atmospheric circulation presented in these pages reveal a considerable variety of patterns.

(1) Secular changes of prevailing strength of the zonal circulation are certainly important, and they are so large that they must surely correspond to considerable variations in the energy effectively available in the troposphere. An example is the increase of zonal circulation illustrated in January from the early 1800s to around 1930.

(2) Both seasonal and secular changes of latitude of the prevailing westerlies tend to accompany shifts of the mean limit of snow- and ice-covered surface, but these sometimes occur independently or precede shifts of this limit.

(3) Displacements of center (eccentricity) of the northern hemisphere circulation can be identified in summer, when they are common, and probably occur in winter also, as is certainly the case in the southern hemisphere. Eccentricity of the circumpolar whirl is probably involved in some of the cases noted, when climatic shifts toward colder or milder winters occur in antiphase in the Pacific and Atlantic sectors, and must have been a pronounced, quasi-permanent feature of the Quaternary ice ages.

(4) Shifts of preferred longitude for cold troughs and warm ridges and changes of wave length in the upper westerlies are also important, and tend to show the expected relationship to changes of mean strength of the zonal westerlies. Corresponding changes in the prevailing longitudinal extent (scale) of anticyclones over Europe may be detected.

(5) There are suggestions of periodic oscillations in the strength of the westerlies and in the frequency of blocking on time scales ranging from days to a half century or more.

When the zonal westerlies are weak, the circulation patterns seem particularly varied (at least as regards preferred longitudes for northerly outbreaks in the east Atlantic and European sectors), possibly depending upon just how weak and how far south or north the main westerlies are. The evidence suggests occasional decades in which easterly winds are prominent in European winters both in epochs of generally strong tendency for westerlies and in epochs when the westerlies are comparatively weak and much less prevalent.

Physical causes must be sought for changes of circulation vigor, since 1100 A.D., of the magnitude indicated in these studies. A close association between atmospheric circulation and prevailing surface-water temperatures in the northern Atlantic has been demonstrated, although the distribution of warm or cold water and ice seems rather to *follow* the changes in the winds. Changes of sea level over the centuries examined can hardly have been sufficient to affect significantly the volume of water transport over shelves and submarine ridges. The most plausible remaining causes suggested for the changes over this period involve the circulation of the deep Atlantic and Pacific oceans and

possible solar and volcanic effects. All these require much further investigation, using where possible circulation maps and measures of circulation strength. The dust from an individual great eruption would take from 2 to 3 years to settle substantially out of the atmosphere. Examination of the extent of the dust veils and postulated effects on world temperature, the amount of polar ice, and the atmospheric circulation strength is planned. It is tempting to attribute to some considerable extent the cold winter climate of much of the northern hemisphere between 1550 and 1850, and the other phenomena of the Little Ice Age culminating about that time, to frequent volcanic dust veils partially screening the solar radiation and intercepting it in the stratosphere while allowing the exit of terrestrial long-wave radiation largely unhindered ("reverse greenhouse effect").

Auer (1958, 1960) has drawn attention to the occurrence of several apparently more or less world-wide waves of volcanic activity in late glacial and postglacial times, notably (**I**) broadly around 7000 B.C., (**II**) broadly around 3000 B.C., (**III**) between 500 and 0, B.C., and (**IV**) since 1500 A.D. (A series of earlier eruptions in the late glacial period seem to have been associated with a great rise of sea level; if eruptive activity under the antarctic icecap was involved, this is a special case where the effects on world climate should be different and attributed chiefly to changes in the ocean circulation rather than to any temporary volcanic dust veils.) The output of volcanic ash and finer dust was in all these cases very large, but nothing is known as to how far veils of fine dust particles were sustained in the atmosphere over periods of a century or more: except for occasional information in the case of the latest volcanic period **IV**, above, since 1500 A.D. Some if not all of the earlier postglacial waves of volcanic activity may however have been associated with climatic recessions, the evidence being suggestive at least in the case of the wave of period **III** about the time when the widespread Grenz-horizont in European peat bogs indicates the onset of the cool, moist "sub-Atlantic" climate (900 to 300 B.C.). Equatorward shifts of the subpolar depression tracks in both hemispheres appear to be indicated about that time, judged by botanical evidence of rainfall (orographic and lee affects) in Europe and Tierra del Fuego. In the case of period **IV**, since 1500 A.D., however, the climatic zones seem to have been displaced somewhat south in both hemispheres.

The botanical (peat-bog) studies alluded to (Godwin, 1954, 1956; Godwin and Willis, 1959), and a wide range of historical, archeological (Butzer, 1958), and glaciological (Manley, 1959) studies suggest it might be possible to piece together rudimentary maps of prevailing winds, depression tracks, and anticyclonic belts for various periods since the ice age. Attempted reconstructions of the prevailing summer and winter circulation patterns around the maximum phases of the Quaternary ice age first published by Simpson (1934) have been modified and refined by me (Lamb, 1961) in the light of modern understanding of the general atmospheric circulation. Further refinement will probably depend upon more detailed geological and glaciological evidence, possibly of successive phases of the ice age. Similar reconstructions should certainly be attempted of the prevailing atmospheric flow patterns, over periods of about 50 years to a century, marking the culminating point of every im-

portant climatic phase since the ice age. Such maps should throw valuable light on the meteorological problem as well as on the retreat and advance of glaciers and vegetation and the development of the arid zones.

Summary

The scientific purpose of studies of climatic change is to reveal the behavior of the atmospheric circulation and throw some light upon the external controls that affect it.

This problem is here discussed in terms of both seasonal and secular changes of circulation pattern and of the heating patterns from which the energy is derived. Circulation maps for Januarys of the period since 1760 are presented together with examples from other recent months, illustrating results already obtained from these studies in the Meteorological Office in London, England, and their relevance to the problems of the future. Documentary data regarding winter weather in different European longitudes since 1100 A.D. and in Japan since 1440 are submitted to an analysis that suggests the atmospheric circulation changes that have occurred during the last 5 to 8 centuries.

Substantial changes in the prevailing strength of the zonal circulation may be deduced, as well as shifts in the longitudes most affected by northerly and southerly winds at different times, and some decades of pronounced easterly winters in temperate Europe. Oscillatory changes have possibly played a minor role, and oscillations of periods ranging from days up to one-half century or more may be suspected. Interactions with the ocean circulation are more clearly demonstrated, although the ocean changes appear rather to follow those in the atmosphere. Changes of latitude of the main zonal westerlies in the Atlantic sector are sometimes associated with eccentricity of the circumpolar vortex. At other times it appears that strong development of the circulation on either the Atlantic or Pacific side of the hemisphere may interfere with that on the other side. This paper concludes with a brief preliminary survey of some possible physical causes of changes in the energy available to the circulation, most attention being given to volcanic dust.

A plea is made for attempts to establish at least rudimentary pictures of the prevailing wind patterns in all the more important climatic phases of post-glacial times.

Acknowledgments

I am indebted to D. J. Schove of St. David's College, Beckenham, Kent, England, for making available his as yet unpublished compilation of records for the southeastern half of Great Britain extracted (and critically examined for consistency) from manorial accounts, tax records, and other sources. My thanks are also due to the Director General of the Meteorological Office, London.

References

Angot, A. 1895. Premier catalogue des observations météorologiques faites en France depuis l'origine jusqu'en 1850. Ann. Bureau central météorol. France, Paris.

Arakawa, H. 1954. Five centuries of freezing dates of Lake Suwa in central Japan. Arch. Meteorol. Geophys. Biol. **B6:** 152–166.

AUER, V. 1958. The Pleistocene of Fuego-Patagonia. Part II. Helsinki (Suomen Tiedeakatemia).

AUER, V. 1960. The Quaternary history of Fuego-Patagonia. Proc. Roy. Soc. London. **B152:** 507–516; 533–538.

BETIN, V. V. 1957. Ice Conditions in the Baltic and Its Approaches and Their Long-term Variations. **41:** 54–125. Trudy. Moscow, USSR.

BROOKS, C. E. P. 1949. Climate Through the Ages. 2nd ed. London, England.

BUCHINSKIY, I. E. 1957. Past Climate of the Russian Plain. Leningrad, USSR.

BULL, E. 1916. Islands klima i oldtiden. Geogr. Tidskrift. **23:** 1–5. Copenhagen, Denmark.

BUTZER, K. W. 1958. Studien zum vor- und frühgeschichtlichen Landschaftswandel im Sahara. Mainz (Akad. der Wiss. und Lit., Abhandl. Math.-Naturwiss. Klasse Nr. I).

EASTON, C. 1928. Les hivers dans l'Europe occidentale. Leyden.

GODWIN, H. 1954. Recurrence surfaces (Grenzhorizonte). Danmarks Geologiske Undersögelser, II Raekke, Nr. 80. Copenhagen, Denmark.

GODWIN, H. 1956. History of the British Flora. Univ. Press., Cambridge, England.

GODWIN, H. & E. H. WILLIS. 1959. Radiocarbon dating of pre-historic wooden trackways. Nature. **184:** 490–491.

HENNIG, R. 1904. Katalog bemerkenswerter Witterungsereignisse von der ältesten Zeiten bis zum Jahre 1800. Berlin. (Abhandl. des kgl. Preuss. Meteorol. Inst. Bd. II, No. 4.)

KRAUS, E. B. 1960. Synoptic and dynamic aspects of climatic change. Quart. J. Roy. Meteorol. Soc. **86:** 1–15.

LAMB, H. H. 1957. Research in world weather patterns. Marine Observer. **27:** 101–110.

LAMB, H. H. 1958. Differences in the meteorology of the northern and southern polar regions. Meteorol. Mag. **87:** 364–379.

LAMB, H. H. 1961. Fundamentals of Climate. Chap. II. *In* Palaeoclimates. Interscience. New York, N.Y.

LAMB, H. H. & A. I. JOHNSON. 1959. Climatic variation and observed changes in the general circulation. Geogr. Ann. **41:** 94–134.

LAMB, H. H. & A. I. JOHNSON. 1960. The use of monthly mean "Climat" charts for the study of large scale weather patterns and their seasonal development. Weather. **15:** 83–91.

MANLEY, G. 1959. The late-glacial climate of N. W. England. Liverpool and Manchester Geol. J. **2** (part 2): 188–215.

SIMPSON, G. C. 1934. World climate during the Quaternary period. Quart. J. Roy. Meteorol. Soc. **60:** 425–478.

VANDERLINDEN, E. 1924. Chronique des événements météorologiques en Belgique jusqu'en 1834. Académie Royale. Brussels, Belgium.

LATE AND POSTGLACIAL CLIMATIC FLUCTUATIONS AND THEIR RELATIONSHIP TO THOSE SHOWN BY THE INSTRUMENTAL RECORD OF THE PAST 300 YEARS

Gordon Manley
Department of Geography, Bedford College, University of London, London, England

I propose in this paper to compare the evidence we have with regard to the characteristics of the smaller climatic fluctuations of recent centuries with that forthcoming with regard to late and postglacial variations. Most of this evidence relates to the temperate lands on either side of the North Atlantic. If we are to study climatic fluctuations from a quantitative standpoint, they, or or their effects, are principally epitomized by the variations of the mean temperature. We may use these as a guide to the probable changes in the surface atmospheric circulation, combined with the changes in the surface temperature of the ocean.

To my mind the most significant feature of the large-scale climatic changes comprehended by the last Ice Age is that the vast accumulation and extension of ice sheets in the Northern Hemisphere that probably began in Greenland in the Miocene (Wager, 1933), after several million years became so markedly eccentric with respect to the Pole; and that it took place on either side of the North Atlantic. This stands in marked contrast to the events in the Southern Hemisphere. Indeed, most of the enlargement of the ice cover in other parts of the world can be ascribed to the spread of mountain glaciers, as for example in the Cordilleran region of North and South America. It is the existence of the wide North Atlantic that differentiates the Northern Hemisphere; but we might well consider the mechanism of its cooling, both on the surface and at depth (Manley, 1950).

The prominent climatic amelioration of the past few decades and the consequent diminution of the extent of the Arctic sea ice have been associated with increased meridional transport of warmth into the Arctic Basin, mainly through the medium of the atmosphere, but partly through the medium of the surface waters of the ocean acting as a reinforcement of the atmospheric circulation.

It is to be noted, following Ahlmann, that the effects of the recent climatic amelioration have been most marked around the northern Atlantic Ocean. Correspondingly, if there is to be a recession, we should likewise expect its effects to be most marked in that region. This leads to the conclusion that, if climate is temporarily unstable, it is most unstable in that part of the world where the meridional transport of heat is most likely to be complicated by the additional effects arising from the presence of an ocean in part of which a surface ice cover is found. It is the presence of the wide oceanic channel between Greenland and Scandinavia that provides a primary element affecting the location and amplitude of climatic change. This becomes evident from the characteristics of the Pleistocene extension of the ice, from the fact that the minor development of glaciation represented by the Post-Alleröd recession was approximately contemporaneous on either side of the North Atlantic, and from the distinct association between the temperature fluctuations shown on either side of the North Atlantic (Manley, 1955).

For the moment let us adopt the hypothesis that there exist variations in the character of the atmospheric circulation over the Northern Hemisphere that from time to time can be described as "meridional" or "zonal" to a greater or less extent. These variations we might attribute either to changes in the output of solar energy, or in the amount of solar energy received at the earth's surface. However if, with a more zonal type of circulation, there is a tendency for the Icelandic Low to be displaced eastward, we should presume that the northwest Atlantic would become cooler. More ice coming down in the East Greenland current and wider spread of the cooler surface waters, perhaps accompanied by greater evaporation into the overlying colder and drier airstream, should be a consequence. Such a cooling of the waters of the northwestern Atlantic should affect the temperatures of all the countries of Western Europe, especially in the spring months when we might expect the effects of ice to be more marked. Moreover, such a narrowing of the belt of warmer water off the European coast should itself tend to cause the initial displacement of the circulation to persist.

Relationships of this kind have long ago been repeatedly examined (for example by C. E. P. Brooks) without any very satisfactory results in individual years on account of the number of variables involved. Schell's work on the Labrador ice, described elsewhere in these pages, should also be recalled. But the length of the available instrumental record in Western Europe is now sufficient to enable us to think in terms of effects spread over decades rather than single years, and it is on some of these that I propose to comment; in this connection I shall discuss recent findings with regard to the magnitude of the fluctuations during the Late and Postglacial.

Let us first consider the noteworthy Late-Glacial climatic oscillation, that is, a warmer phase followed by a recession that first became known in the Alleröd deposits. There was an earlier Bölling oscillation but we do not yet know enough about it. The Post-Alleröd climatic recession, we now know, prevailed for about five or six centuries in the ninth millenium B.C. The average of a number of radiocarbon datings puts the limits at 8860–8200 B.C. The evidence of the Windermere lake sediments shows that, among the mountains of northwest England, glaciers had previously entirely disappeared. Varved clays, overlying the earlier detritus silt containing some tree pollen, testify to the re-establishment of glaciers within the Windermere basin, and numerous morainic deposits bear witness to their extent. Assemblage of the evidence from the very wet central valleys of the Lake District and the drier regions to the northeast and east goes to show that the snow line rose towards the drier regions by an amount that is in agreement with a distribution of precipitation like that of the present, declining towards the north and east in the same ratio. From this it may further be deduced that the water equivalent of the annual total precipitation was very much the same as that prevailing today (Manley, 1959). It appears evident that a similar arrangement developed in North Wales.

However the fall of the snow line necessary for the re-establishment and maintenance of these glaciers is such that the onset of the climatic recession must be deduced to have been marked by a fall in the mean summer temperature of 4° to 5° C. or 7° to 9° F. The volume of the glaciers is such that, making

reasonable assumptions, their re-establishment and extension to their known limits would require about 80 years of prevailingly disturbed weather, that is, the predominance of a "zonal" type of circulation with its characteristically cloudy unsettled summers and stormy winters. Approximately 300 years, then, appears to have been required for the glaciers gradually to melt and disappear. That the time required for retreat should be three or four times as long as that required for advance is not out of keeping with what we know of the oscillations of the past four centuries.

However the question now arises, how would all this happen? It is customary to regard glacier growth and advance beside the North Atlantic as likely to be a concomitant of a "zonal" type of circulation. If we make use of the instrumental data representative of central England, the difference between a favorable "warm-anticyclonic" type of summer and an "unsettled westerly" summer is commonly of the order of 3° C. even within the same decade, as with 1959 and 1954 (2.6); 1922 and 1921 (2.5); 1899 and 1890 (2.9); 1868 and 1860 (3.4); 1826 and 1823 (4.0). Summer is taken as the three months, June to August.

The illustrations here given (FIGURES 1 and 2) will show the kind of distribution of open water and drift ice that would be likely to be found in the Late-Glacial Atlantic. In such circumstances a change from "fair anticyclonic" to "cool westerly" conditions in summer would be accompanied by a greater fall of temperature than today. At Stykkisholm in west Iceland, which lies closer to the east Greenland drift ice, the present-day range of temperature between good and bad summers is about one half as great again as in England.

Hence, with an icy North Atlantic, a fall of the mean summer temperature of 4° to 5° C. can readily be interpreted as the result of a simple change of circulation type, accompanied by an eastward displacement of the Icelandic Low such as we might expect today. The question remains, however, what mechanism must be postulated in order that the circulation can remain persistently "zonal" for several decades.

Let us now turn to the postglacial fluctuations. Iversen's evidence (1944) from studies of the occurrence of the pollen of ivy, holly, and mistletoe in Denmark sets limits to the range of variation of temperature well within the capacity of our present-day atmospheric circulation, in association with slight changes of sea-surface temperature. The climatic deterioration at the onset of the Sub-Atlantic is well known. It implies a fall in the mean summer temperature of about 2° C.; but over how long a period was it necessary for the summer temperature to fall and the rainfall to increase, accompanied by a decline in the evaporation sufficient to cause a marked change in the character of the forests and a fall in the altitude of the tree-line? This we do not know, but from conversations with botanists I gather that a period of 50 to 100 years might be acceptable.

Furthermore the more we learn about the details of the postglacial the more it appears that there were minor fluctuations sufficient to leave their evidence in some regions but not in others.

With regard to historic time, we have the evidence from Butzer's studies in the Near East and North Africa (1957) that there were a number of minor fluctuations of note, for example the "dry period" A.D. 590 to 640. We are re-

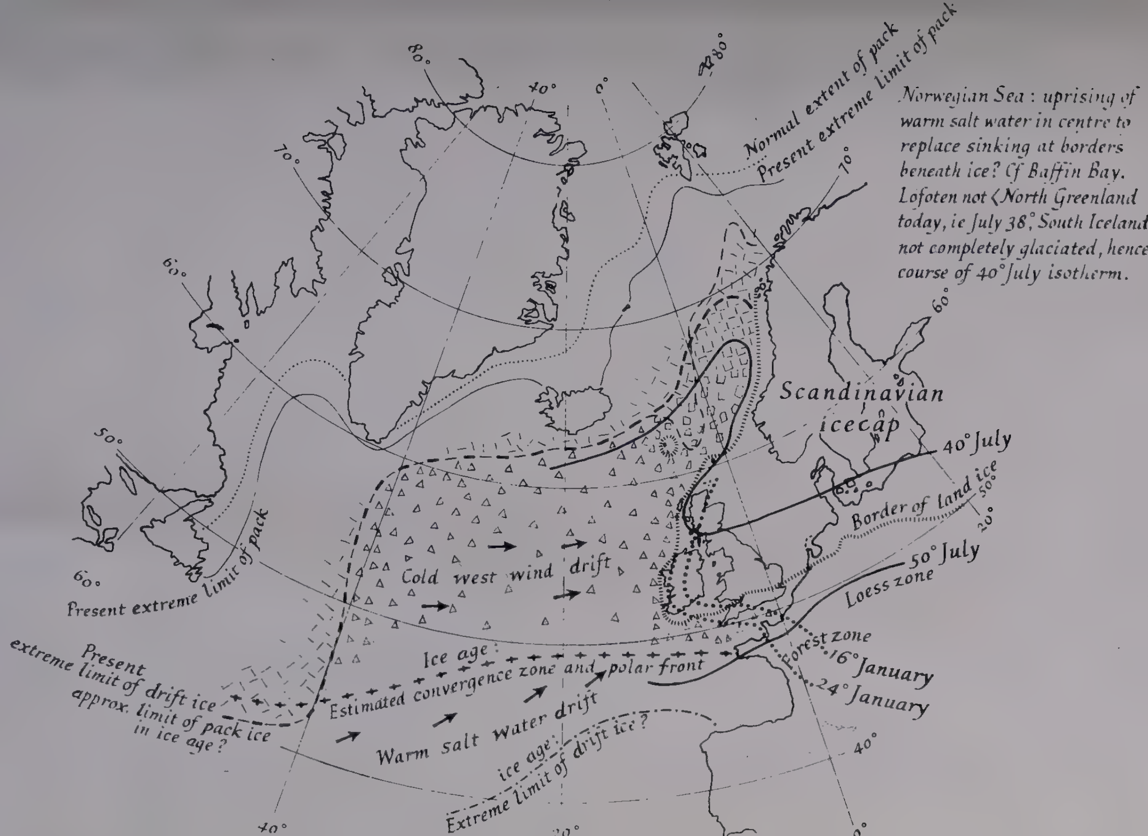


FIGURE 1. The relationship of surface features and the glacial climate of the North Atlantic Ocean. The polar front would often lie about 50° N near the convergence. An Arctic front would probably lie through the Norwegian Sea from south of Iceland. Reproduced by permission of the Royal Geographical Society (Manley, 1951).

minded again of the considerable body of material assembled by Brückner (1890) that led him to conclude that in Europe short-term fluctuations between warmer and colder, wetter and drier groups of years had prevailed for many centuries past. Whatever view is now taken of Brückner's work we can safely accept the fact that groups of years with a "prevailing tendency" can be recog-

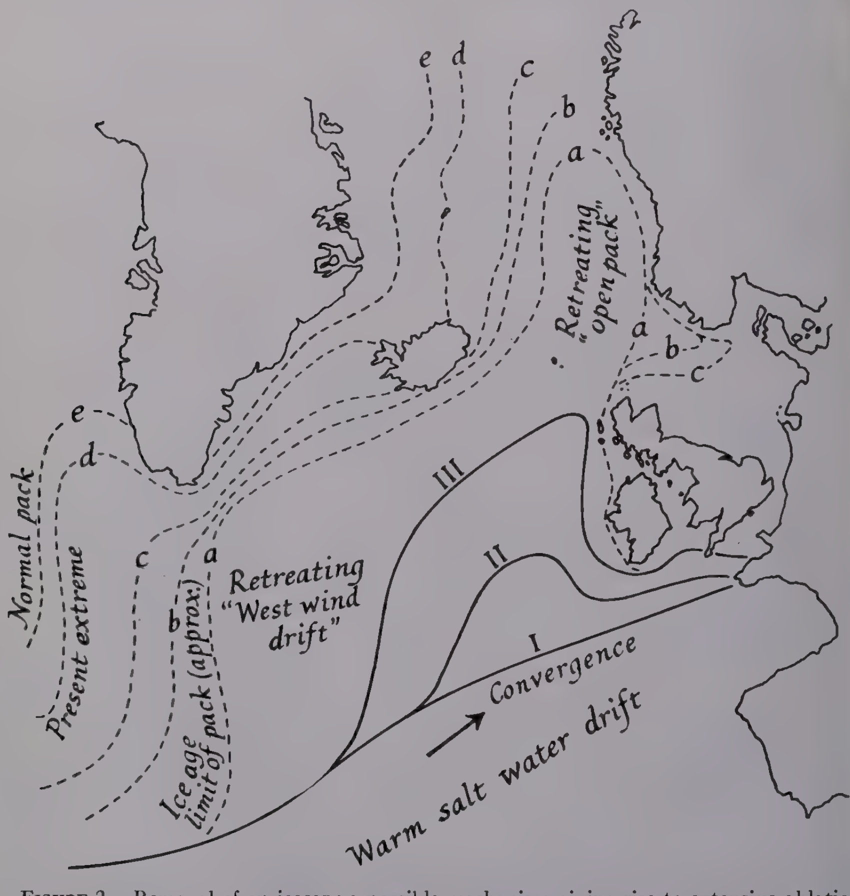


FIGURE 2. Removal of an icecap: a possible mechanism giving rise to extensive ablation by convection from a warmer sea. III: a hypothetical intermediate stage corresponding roughly to "Perth Readvance." Reproduced by permission of the Royal Geographical Society (Manley, 1951).

nized; Brooks (1928) for example was fond of emphasizing the rainy character of the late 11th century. Flohn (1954) has emphasized the unsettled cool period about 1428–1460 and the warmth of the early 16th century. The beginning of a climatic recession leading to the very marked advance of the Alpine glaciers appears to be recognizable about 1550. In the United States we may note Schulman's discussion (1956) of the tree-ring evidence and the dry periods in the Southwest in the 13th century (1215 to 1299) and again farther north in the late 16th century (1573 to 1593). However in Europe we may also note the

rather controversial evidence with regard to the extent of the "climatic recession" about 1300, which left its mark in the Scandinavian pollen record, but which does not appear to be so prominent farther east. At this stage it is appropriate to recall Koch's paper (1945) on the East Greenland ice, and the relatively limited development of ice around the coasts of Iceland in the 14th century compared with the period from 1600 onward.

It becomes evident that every one of these climatic fluctuations around the North Atlantic might be explained in terms of a simple change in the prevailing character of the atmospheric circulation, leading to temperature changes of much the same magnitude, but varying considerably in their duration, or persistence.

Let us now see what the instrumental record has to tell us. Largely on account of Ahlmann's (1953) work on the factors governing the variations in the extent of glaciers, interest since 1940 has been revived in the provision of the longest possible series of monthly mean temperatures. In northwest Europe some of the best known include Labrijn's very careful reduction for Utrecht beginning in 1706. More recently we have had "Lancashire" since 1753 (Manley, 1946), Basle since 1755 (Bider *et al.*, 1959) and Birkeland's revised table for London since 1763 (Birkeland, 1957), to add to the earlier Stockholm, Oxford, Copenhagen, Bergen, and Edinburgh series.

With the aid of the above Oxford and Lancashire records combined with others from the English Midlands, a composite series has been built up representative of "central England" back to 1698, but with a gap 1707 to 1722 that has had to be filled by estimation from the Dutch records, supplemented by English noninstrumental diaries. While almost any temperature record kept before 1800 is only too open to criticism, the mutual support given by these several long-period tables back to 1760 is noteworthy and lends considerable credence to deductions with regard to trends, as Labrijn has shown (1948).

Recently I have made reductions of some very primitive instrumental observations stemming from the pioneers of the Royal Society and, notably, Robert Hooke. These, in combination with the data from some newly-found meteorological journals covering the period 1669 to 1717, enable a series of estimates to be given of the mean temperature for each month, in the London area, back to 1680. Daily observations of the wind direction and the occurrence of rain or snow enable one to discuss the character of the period in some detail.

As these appear to be the earliest instrumental records from N.W. Europe that are sufficiently long to be capable of useful interpretation and an attempt at the estimation of monthly means, a short note on their reduction is not here out of place.

First, from 1699 to 1706 we have the admirably-tabulated thrice-daily observations by William Derham, F. R. S., giving the readings of a north wall thermometer out-of-doors at fixed hours. I have analyzed these in detail, and am indebted to the work of Louise D. Patterson (1951, 1953) on the 17th-century thermometer scales. As we have insufficient information, it was first necessary to deduce the probable characteristics of Derham's exposure by comparing, for each of the 12 months, the mean rise or fall of his shade temperature between the hours of observation, with the averages published for each hour in

the same months at Kew and Greenwich observatories. Second, the probable value of Derham's degree can be reasonably checked against the average and absolute extremes of temperature recorded. Third, the over-all mean temperature of the six winter months can be checked, approximately, against the frequency of snowfall during the period. Fourth, there is a very tenuous overlap of a single year (1706) with Labrijn's tables for Holland. Scrutiny of Derham's record leads to the view that his zero did not significantly change.

Derham's outdoor observations are in turn overlapped by the recently found observations in manuscript (1692 to 1703) kept by John Locke, who kept an indoor thermometer made by Thomas Tompion, in what appears to have been a characteristically drafty English sitting-room. Indoor mean temperatures can be shown to bear a reasonable relationship to those kept outdoors. Unfortunately Locke's series is much interrupted, and for a number of the summer months all one can do is make estimates from other contemporary journals.

From 1680 to 1694 we have a further set of daily observations by John Downes, physician to Christ's Hospital in London. His thermometer was extremely imperfect and, after a break in 1691, he seems to have acquired a new one. Hence we have no satisfactory overlap with the later records; a very scanty series of observations by Locke at various times between 1681 and 1683 does little but provide an indicator of the change of Downes' zero. Hence the monthly means can be assembled only by making assumptions with regard to the unlikelihood of extreme values lying outside those of the past 150 years and then adjusting the values for the whole decade to agree with the over-all frequency of snowfall. The results, however, are consistent with contemporary accounts (see TABLE 1).

Reduction of these very primitive 17th century observations is desirable, as the importance of the period is manifest. First, this period lies within the well-recognized period of feeble sunspot activity, 1645 to 1715. One may gather from a recent table (Schove, 1955) that a particularly weak maximum in 1693 was followed by an exceptionally low minimum in 1698. Second, the 1690s have long been notable for bad harvests, distress, and considerable economic effects in northwest Europe; and a decided advance of Alpine and Icelandic glaciers is known to have set in, ending about 1715. Third, 1695 stands out as the worst "ice year" in the Icelandic Annals; Koch mentions that the sea ice was observed around to the southwest coast. The year 1698 in England is described as giving the latest harvest on record.

Even if some quite serious undetected error exists in the reduction of these very primitive instrumental observations, the fact that emerges at once from the reduction of the available material, especially the snowfall data, is the remarkable magnitude of the short-term climatic recession in this last decade of the 17th century (FIGURE 3). From 1692 to 1701 in particular, cold springs and cool unsettled summers were predominant. Examination of the frequency of precipitation within one homogeneous record, covering 1668 to 1700, shows an 11 per cent increase in the average annual number of days with rain or snow falling in 1691 to 1700 compared with the two preceding decades; and if we take snow alone, a 50 per cent increase in the average number of days with snow to a total twice that expected today in London. This provides us with a reasonable general check—indeed the only check at present beyond descriptive

statements—on the over-all averages of temperature during the colder months. Taking account of the cold unsettled springs in particular it seems to me that we must recognize the combined effects of a strong zonal flow, a marked increase in the input of ice and cold water into the northwestern Atlantic, and probably

TABLE 1*

ESTIMATES OF THE MEAN MONTHLY TEMPERATURE (°F.) IN THE LONDON REGION, 1680 TO 1706

	J	F	M	A	M	J	J	A	S	O	N	D	Year
1680	(41)	(39)	(43)	45	52	57	63	61	60	52	45	34	49.3
1681	34	36	40	48	53	59	61	63	59	54	45	38	49.2
1682	44	37	41	45	55	59	61	60	57	51	43	43	49.7
1683	39	37	43	51	55	63	62	58	57	45	41	33	48.7
1684	27	31	38	45	57	61	63	62	56	53	38	40	47.6
1685	34	39	42	49	56	60	59	60	55	54	46	44	49.8
1686	45	44	46	49	56	62	63	60	57	50	45	43	51.7
1687	39	41	41	45	54	57	62	61	53	53	44	43	49.4
1688	39	36	39	43	53	57	62	60	55	46	40	38	47.3
1689	34	41	42	48	53	56	62	61	57	48	42	41	48.7
1690	40	41	41	48	51	57	62	61	56	49	45	41	49.3
1691	35	35	42	45	52	58	61	62	54	(50)	(42)	39	47.9
1692	37	32	40	47	50	58	61	61	54	45	42	39	47.2
1693	38	42	38	45	50	60	61	61	55	51	44	38	48.6
1694	33	42	39	47	50	57	61	57	52	(47)	43	(37)	47.2
1695	31	33	39	43	50	(57)	58	(57)	(54)	50	(43)	40	46.2
1696	43	41	39	43	53	(57)	(62)	(62)	(55)	(50)	(43)	37	48.7
1697	35	34	43	46	55	57	(62)	(60)	(56)	(50)	(40)	37	47.8
1698	33	34	39	47	49	56	61	(61)	(57)	(50)	40	39	47.2
1699	38.5	39	40.5	45	51.5	60.5	65.5	61	58	50.5	42.5	39	49.3
1700	40	37.5	39.5	45	55.5	58.5	61	61	57.5	49.5	41.5	40	48.9
1701	38	37.5	38	41	53	59.5	67.5	63	60.5	47	44.5	39	49.0
1702	42	45	44	44.5	52.5	58	61	63	60	51.5	41	40.5	50.3
1703	36.5	40	43.5	49	55	59	63	63	52.5	47	46	42	49.7
1704	36.5	38.5	43	49	54	60	64	64.5	55.5	48.5	44.5	39	49.7
1705	37	39.5	41	47.5	53.5	56	62	65.5	55.5	49.5	40	40.5	49.0
1706	37.5	40.5	45	50	55.5	62	63	64	56.5	53.5	44	41	51.0
1681/1690	37.5	38.3	41.3	47.3	54.3	59.3	61.7	60.6	56.2	50.3	42.8	40.4	49.1
1691/1700	36.3	37.0	39.9	45.3	51.5	57.8	61.3	60.4	55.2	49.3	42.1	38.5	47.9
Probable values at Upminster today, based on *rural* stations around London:—													
1921/1950	39.9	43.0	43.8	48.3	54.1	59.8	63.4	62.7	58.5	51.3	44.4	40.4	50.6

Average number of days with rain and days with snow observed to fall:
1921/1950, approximately 168 (0.01″ or more) and 12.5 (Good, but not continuous, observation)

1671/1680, as observed:	155	17.2
1681/1690, " "	167	18.2
1691/1700, " "	179	25.5
1701/1710, " "	168	17.9

* Primarily based on Derham's Upminster record, 1699 to 1706; partially overlapped by Locke's incomplete record, 1692 to 1703. Earlier years less reliable, as there is no satisfactory overlap with the later records. Estimates for this period, and for Locke's missing months (in brackets), have taken into account the daily observations of wind and weather in other contemporary journals, and are given to whole degrees only (1699 onward to 0.5°).

a displacement of the axis of an Atlantic upper trough, leading to persistent instability over Britain. The cool unsettled summer of 1954 gives us a model. It may also be noted that, according to Flohn (1954), the period 1680 to 1740 in Central Europe was for the most part "mild oceanic"; that is, the short-term recession of the 1690s was not prominent in Central Europe, although its effects were very marked in Sweden, the uplands of central France (Etienne, personal communication) and in the Western Alps.

In view of the fact that the over-all fluctuations of the mean summer temperature in northeastern North America show considerable association with those of northwest Europe (Manley, 1955) it might be worthwhile investigat-

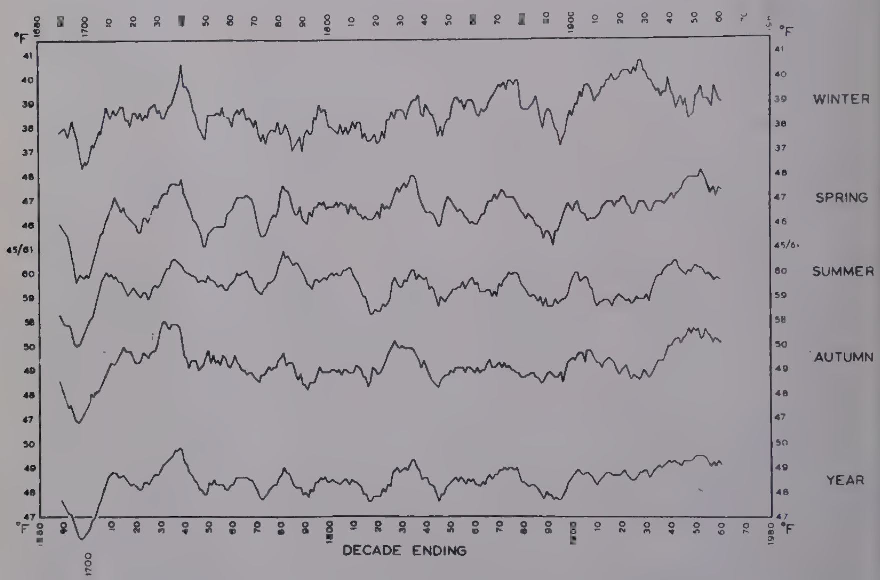


FIGURE 3. Decadal running averages of seasonal and annual mean temperatures for central England: 1680 to 1960.

ing whether there are any records of a similar prevalence of cool seasons in New England between 1688 to 1702 or thereabout.

The unusual character of this decade is such that it may perhaps appear as a "climatic marker" in records in other parts of the world. On the other hand, if we find that it effects are confined to the periphery of the North Atlantic, as the present evidence suggests, we have the more reason to seek for its cause in the injection of cold water consequent upon a temporary change in the character of the circulation, accompanied by a predominance of strong northerly winds on the west side of the Norwegian Sea.

The appearance of the early 18th century temperature records indicates an equally marked preponderance of warmth between about 1723 and 1739. This has of late been the subject of comment among Swedish economic historians; it was notable for mild springs and warm autumns (FIGURE 3).

Inasmuch as there were several notable volcanic eruptions between 1680 and

1693 (including Hekla in 1693), and again especially in Iceland about 1720 to 1727, there seems to be no obvious association between these climatic episodes, as a whole, and vulcanicity. We may however note that both the years 1694 and 1695 were characterized by cool summers, while in the later decades 1725 was outstandingly cool in an otherwise warm period. Hence the possibility exists that volcanic effects were superimposed on those arising from other causes.

There are of course several later groups of years during which cool unsettled summers have prevailed, for example 1809 to 1816 and 1879 to 1886. But it is not possible to find any very prolonged series of such years. Hence we are impelled to ask why the oscillations of temperature in the past 300 years appear to have been more marked, but shorter-lasting, than previously. (In regard to winter temperatures, it is somewhat noteworthy how often exceptionally mild seasons have followed exceptionally cold seasons in the same decade: for example, 1684 and 1686; 1795 and 1796; 1916 and 1917; 1942 and 1943.)

Here we may look to Koch's paper on the East Greenland Ice, which appears to show "pulsations" of some magnitude since 1600. We may also note Schulman's remarks (op. cit., pp. 56, 58, 69) based on his studies of tree rings, that the wave length of the fluctuations they display tends to be shorter after 1630 or thereabouts, than during the previous century. Third, the marked development of glaciation over much of Europe and North America has taken place since 1550.

All this can be taken to suggest that around the North Atlantic, at least, the climate has become more "unstable" during the past 3 or 4 centuries, accompanied by decided oscillations of the spring and summer mean temperature, which in turn have affected the glaciers wherever the Atlantic maritime-polar air has penetrated. This can be shown if we observe the association between the small fluctuations of the spring temperature in England and the behavior of the glaciers of the Western Alps. It is noteworthy that advance has not begun yet despite the first signs of decline of the spring temperature; to all appearance (FIGURE 3), the ten-year average should fall below 47° F. or 8.3° C. before advance can be expected. Here is where we may ask: By what stages and when will the northwestern Atlantic become cooler? What mechanism exists within the system "atmosphere-ocean" that can lead to a longer-term persistence of favorable or unfavorable summers?

In conclusion: first, the evidence demands "persistence of circulation patterns," in the past, over periods that may be many decades in length. Second: the North Atlantic climate appears to have become rather more unstable, with shorter-term fluctuations, during the past 3 or 4 centuries; perhaps this is attributable to the secondary effects arising from the vicissitudes of the Arctic ice. Third: such evidence as we have for the marked "oscillation" adjacent to the northeast Atlantic between 1688 and 1739, in which a very cold decade and a very warm decade were found, does not suggest any close association with either vulcanicity or sunspots. Accordingly there would appear to be a case for more attention to the factors leading to variations in the transport of heat by the water flowing into and out of the Arctic Basin. Before we can look into other matters we should explain this question of varying persistence of fluctuations; especially we might seek for an explanation for what we may

call long- and short-wave patterns of climatic oscillation. I suggest that it may lie with the oceanographers to take up this challenge and to take up, for example, the question of how great the accumulation of snow must be within the Arctic Basin in order that the surface efflux of relatively fresh water will again prevail throughout a wider area of the northwest Atlantic.

It seems to me that we should develop further the implications of what we may call Faegri's palaeoclimatological principle (Faegri, 1950, p. 192): "the shorter the duration of a climatic fluctuation, the smaller is the area similarly affected. The longer the cycle, the greater the area within which it is felt in the same way." It may be that we are approaching the stage when we shall be able to examine this question of persistence over wide areas.

Acknowledgments

I express my grateful acknowledgments to the Officers of the Royal Society, and of the British Museum, in London, England, and of the Bodleian Library at Oxford, for the opportunity to work on early meteorological manuscripts in their keeping.

References

AHLMANN, H. W. 1953. Glacier variations and climatic fluctuations (Bowman Lecture). New York, N. Y. American Geographical Society.

BIDER, M., M. SCHUEPP & H. VON RUDLOFF. 1959. Die Reduktion der 200-jährigen Basler Temperaturreihe. Arkhiv Meteorol., Geophys. und Bioklim. Bd. **9**.

BIRKELAND, B. J. 1957. Homogenisierung der Temperaturreihe Greenwich 1763–1840. Geofysiske Publikasjoner. (Norske Vidensk. Akad.). **XX:** 1.

BROOKS, C. E. P. & J. GLASSPOOLE. 1928. British floods and droughts. London: Benn.

BRÜCKNER, E. 1890. Klimaschwankungen . . . seit 1700. Vienna, Austria.

BUTZER, K. W. 1957. Late-glacial and post-glacial climatic variation in the Near East. Erdkunde Bd. **XI**(1)**:** 21–35.

FAEGRI, K. 1950. On the value of paleoclimatological evidence. Centenary Proceedings Volume, Roy. Meteorol. Soc. London.

FLOHN, H. 1954. Witterung und Klima in Mitteleuropa. Forschungen zur Deutschen Landeskunde, Bd. **78**. . Hirzel. Stuttgart, Germany.

IVERSEN, J. 1944. Viscum, Hedera and Ilex as Climatic Indicators. Geol. Fören. Stockholm Förh. **66:** 463.

KOCH, L. 1945. The East Greenland Ice. Medd. om Grønland. **130**.

LABRIJN, A. 1948. Klimatschommelingen in het stroomgebied van de Rijn. De Watervoorziening van Amsterdam.

MANLEY, G. 1946. Temperature trend in Lancashire. Quart. J. Roy. Meteorol. Soc. **72**.

MANLEY, G. 1950. Glacier variation and climatic fluctuations in Britain. J. Glaciol. **1:** 352–356.

MANLEY, G. 1951. The range and variation of the British climate. Geogr. J. **117**(1)**:** 43–68.

MANLEY, G. 1955. A climatological survey of the retreat of the Laurentide Ice-Sheet. Am. J. Sci. **253:** 256–273.

MANLEY, G. 1959. The Late-glacial climate of N.W. England. Liverpool and Manchester Geol. J. **2:** part 2, : 188–215.

MANLEY, G. 1959. Temperature trends in England: 1698–1957. Arkhiv Meteorol., Geophys. und Bioklim. Bd. **9**.

PATTERSON, L. D. 1951. Thermometers of the Royal Society, 1663–1768. Am. J. Phys. **19:** 523–535.

PATTERSON, L. D. 1953. The Royal Society's Standard Thermometer, 1663–1709. Isis, **44:** 51–64.

SCHOVE, D. J. 1955. The sunspot cycle. J. Geophys. Research. **60:** 127–146.

SCHULMAN, E. 1956. Dendroclimatic changes in semi-arid America. Univ. Ariz. Press. Tucson, Ariz.

WAGER, L. W. 1933. The form and age of the Greenland Ice Cap. Geol. Mag. **70:** 145–56.

SOME STATISTICAL ASPECTS OF LONG-TERM FLUCTUATIONS IN SOLAR AND ATMOSPHERIC PHENOMENA

G. W. Brier

United States Weather Bureau, Washington, D.C.

This paper is a progress report, presenting the preliminary results of an investigation of a number of meteorological time series dating as far back as 1750. These series are critically examined for any evidence of influence by the well-known solar cycle of about 11 yr. by the technique of power spectrum analysis. In addition the data are treated by means of specially designed band-pass filters to recover phase information and to study possible nonlinear effects.

Introduction

Quantitative observations of the solar disc have been made for more than 250 yr., and numerous compilations have been made of the Zurich daily relative sunspot number R first introduced by Wolf about 1849. Recently Chernosky and Hagan[1] have given the monthly and annual values for the years 1749 to 1957, and also a set of annual values for the period 1700 to 1749. The sunspot numbers have been used by many investigators in attempting to determine whether a real relationship exists between solar activity and meteorological variables such as precipitation, pressure, and temperature. In general the results have been inconclusive or controversial, due in part to the crudeness or lack of validity of the statistical methods used and, in part, to the shortness of the meteorological time series available. Very few weather records extend back more than 100 yr., and some of these are not homogeneous. Furthermore even 100 yr. of record involves only 9 solar cycles of 11 yr. and less than 5 cycles of 22 yr. Such a length of record is quite inadequate for detecting small cyclical effects with any degree of confidence.

However in the last two decades the development of some new mathematical techniques for the statistical analysis of time series, along with the rapid progress in the development and use of high-speed electronic computers, makes it practical to re-examine some of the solar and weather data with the hope of obtaining conclusions more definite than many of those obtained previously. This paper reports on the progress of a study applying these modern statistical techniques to some of the longer weather records. Since the data processing is not finished, only preliminary results will be presented here, and final conclusions must await the completion of the statistical analysis. These will be reported in a later paper.

Method of Analysis

Although mathematicians have been familiar with harmonic analysis for centuries, and investigators of empirical phenomena have used some form of periodicity or spectral analysis for many years, it was not until the middle of this century that Tukey[2] suggested a sound and practical computational procedure for obtaining spectral estimates based on the results of the pioneer work

by Wiener[3,4] on generalized harmonic analysis. The recommended procedure provides spectral estimates (U_h) showing how the variance of the time series is distributed as a function of frequency. Ward[5] has recently described the method in some detail in connection with an application to the geomagnetic disturbance indices. Some other applications are discussed by Panofsky and Brier,[6] and a more complete treatment of the subject is given in a recent monograph by Blackman and Tukey.[7]

In this project the spectral estimates were obtained by first computing the sample autocorrelation function

$$R_k = \frac{1}{N-k} \sum_{i=0}^{N-k} x_i x_{i+k}$$

where N is the number of observations used and the x_i are data points expressed in terms of the deviation from the mean of the series. From the Fourier-transform of the R_k function the apparent "line powers" L_h are determined by

$$L_0 = \frac{1}{2M}(R_0 + R_M) + \frac{1}{M} \sum_{k=0}^{M-1} R_k$$

$$L_h = \frac{1}{M}(R_0 + [-1]^h R_M) + \frac{2}{M} \sum_{k=1}^{M-1} R_k \cos \pi k \frac{h}{M}$$

$$L_M = \frac{1}{2M}(R_0 + [-1]^M R_M) + \frac{1}{M} \sum_{k=1}^{M-1} (-1)^k R_k.$$

M is the number of lags for which the autocorrelation function is computed and is usually about 5 or 10 per cent of the number of observations N.

The values of L_h are smoothed to obtain the spectral estimates

$$U_0 = 0.54L_0 + 0.46L_1$$

$$U_h = 0.54L_h + 0.23(L_{h-1} + L_{h+1})$$

$$U_M = 0.54L_M + 0.46L_{M-1}$$

Although power spectrum analysis has been found to be a valuable tool in the study of geophysical time series, it has its limitations and often should be supplemented by other types of analysis. If the system being studied is not nearly linear then, as Tukey[2] points out, "this type of analysis will miss important parts of the story." Also spectrum analysis discards phase information as well as the details of any amplitude variation. Sometimes it may be desirable to recover this information to gain a little insight as to what possibly might be going on in the original series. One way of accomplishing this is by means of band-pass filtering. In this technique the original series X_t is operated on by a "filtering function," or perhaps several such functions. These methods have been discussed by Holloway,[8] Panofsky and Brier,[6] and others. The purpose of such filtering is to eliminate or reduce in amplitude the fluctuations (whether of higher or lower frequency) other than those of a particular time scale under investigation. The procedure is to treat the observations X_t in the time series by the following linear equation

$$F_t = \sum_{k=-n}^{m} W_k \ X_{t+k}$$

where W_k is a particular weight in the filtering function. The weight W_0 is known as the principal weight or the central weight when the filter is symmetrical with $n = m$. In the process of filtering the time series, successive observations are cumulatively multiplied by these weights producing a new series beginning F_t, F_{t+1}, F_{t+2} and continuing in this succession.

The particular set of weights used in this study is listed in TABLE 1 and is shown in graphical form in FIGURE 1. This filter function has been designed

TABLE 1
SET OF WEIGHTS USED IN THIS STUDY

W_{-21}	−0.001	W_0	0.136
W_{-20}	−0.004	W_1	0.107
W_{-19}	−0.006	W_2	0.034
W_{-18}	−0.009	W_3	−0.046
W_{-17}	−0.008	W_4	−0.095
W_{-16}	−0.006	W_5	−0.098
W_{-15}	−0.001	W_6	−0.062
W_{-14}	0.003	W_7	−0.014
W_{-13}	0.007	W_8	0.023
W_{-12}	0.013	W_9	0.038
W_{-11}	0.023	W_{10}	0.034
W_{-10}	0.034	W_{11}	0.023
W_{-9}	0.038	W_{12}	0.013
W_{-8}	0.023	W_{13}	0.007
W_{-7}	−0.014	W_{14}	0.003
W_{-6}	−0.062	W_{15}	−0.001
W_{-5}	−0.098	W_{16}	−0.006
W_{-4}	−0.095	W_{17}	−0.008
W_{-3}	−0.046	W_{18}	−0.009
W_{-2}	0.034	W_{19}	−0.006
W_{-1}	0.107	W_{20}	−0.004
		W_{21}	−0.001

to have a particular frequency response. The frequency response R_f of a filter is a function of frequency and is given by the formula

$$R_f = W_0 + 2 \sum_{k=1}^{n} W_k \cos 2\,\pi\, fk$$

where f is expressed in terms of cycles per data interval and ranges from 0 to ½. The actual frequency response for the filter function used here is shown by the curve in FIGURE 2. The ordinate of this curve gives the ratio of the amplitude of a wave of a given frequency f in the time series after filtering to the original amplitude before filtering. The weights W_k used were determined by the formula

$$W_k = R(0) + 2 \sum_{f=1/2n}^{n/2n} R(f) \cos 2\pi fk$$

$$f = \frac{1}{2n}, \frac{2}{2n}, \frac{3}{2n}, \cdots \frac{1}{2}$$

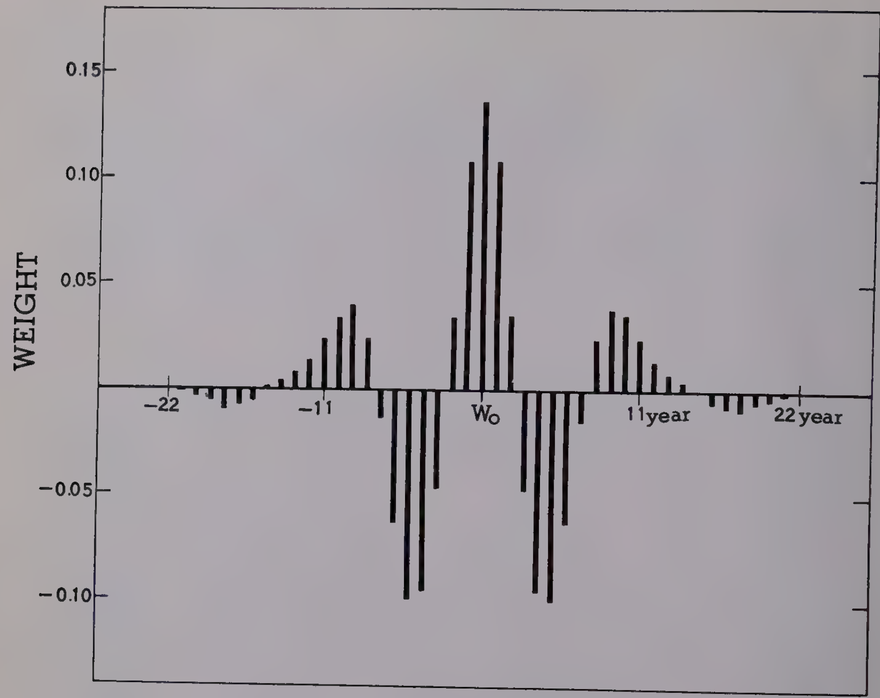


FIGURE 1. Graph of weights used in the 11-yr. filter function.

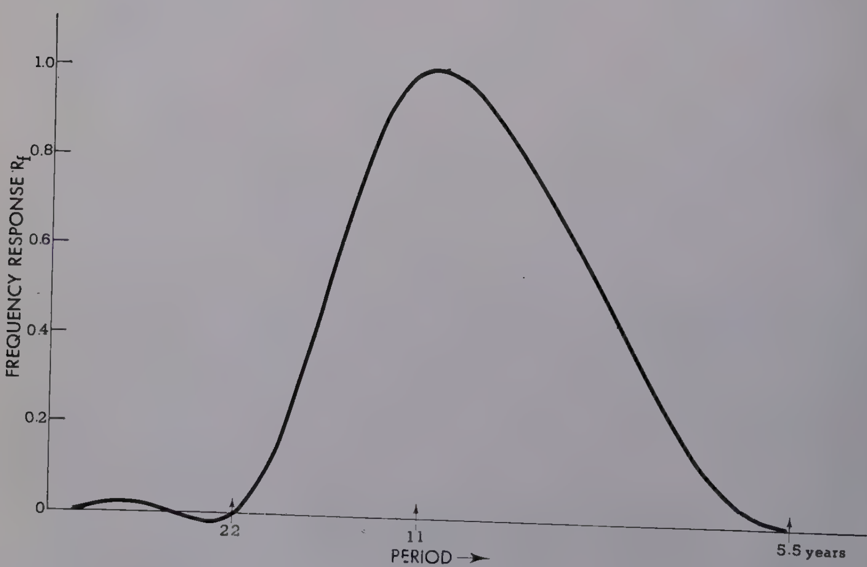


FIGURE 2. Frequency response of the 11-yr. filter function weights.

where $R(f)$ is the theoretical or desired frequency response specified in advance and $-21 \leqslant K \leqslant 21$. For $K > 21$ the weights W_k approach zero very rapidly and remain so small that they can be neglected. The function $R(f)$ was specified to have unit response at a period of 11 yr. ($f = 1/11$) decreasing to zero response at 22 yr. ($f = 1/22$) and at 5.5 yr. ($f = 2/11$). The mathematical functions chosen to represent this were:

$$\begin{array}{ll} R(f) = 0 & 0 < f < 1/22 \\ R(f) = 1/2 + 1/2 \cos 22\pi f & 1/22 \leqslant f < 1/11 \\ R(f) = 1/2 - 1/2 \cos 11\pi f & 1/11 \leqslant f < 2/11 \\ R(f) = 0 & 2/11 \leqslant f < 1/2, \end{array}$$

where $f = 1/2$ is the Nyquist frequency or 1/2 cycle per data interval. It is seen that the curve in FIGURE 2 approaches this function very closely, which means

TABLE 2

STATIONS FOR WHICH AN ANALYSIS OF THE RECORD HAS BEEN UNDERTAKEN

	Length of record used (years)		
Station	Precipitation	Pressure	Temperature
Albany, N. Y.	120*		
Budapest, Hungary	100*	130	168
Copenhagen, Denmark	110*		
Frankfurt, West Germany	110*		
Greenwich, England	110*		
Helsinki, Finland	100*		
Malta	100		
Milan, Italy	160*		
New Haven, Conn.			154
Rome, Italy	140*		105
Stykkisholmur, Iceland		100	

* Reported upon in this paper.

that practically nothing has been lost by truncating the weights at $K = 21$ which is approximately twice the period of maximum response (11 yr).

Results and Discussion

The techniques discussed above have been applied to the weather records indicated by TABLE 2. Only the annual means of the data have been treated, and only the results of the computations for precipitation are completed at this time. Additional results for pressure and temperature will be reported upon later, as well as a treatment of the data using 22- and 5.5-yr. band-pass filters. The choice of the stations used was dependent upon readily available compilations of data such as those published in the *World Weather Records*.[9,10] Consideration also was given to other projects in spectral analysis that are known to be under way in an effort to avoid duplication.

The results of the spectral analysis are not presented here in detail, since no striking or consistent features were noticed. Typical results, such as those for Copenhagen, Denmark, and Greenwich, England, are shown in FIGURES 3 and

4. A composite spectrum found by averaging the results for the 8 stations is shown in FIGURE 5. There is no evidence of spectral peaks at 5.5, 11, or 22 yr., or at any other period.

The result of using the 11-yr. band-pass filter on the precipitation data for Frankfurt, West Germany, is shown in FIGURE 6. Both the original X_t series and the filtered series F_t are shown here. Although the filtered series F_t has

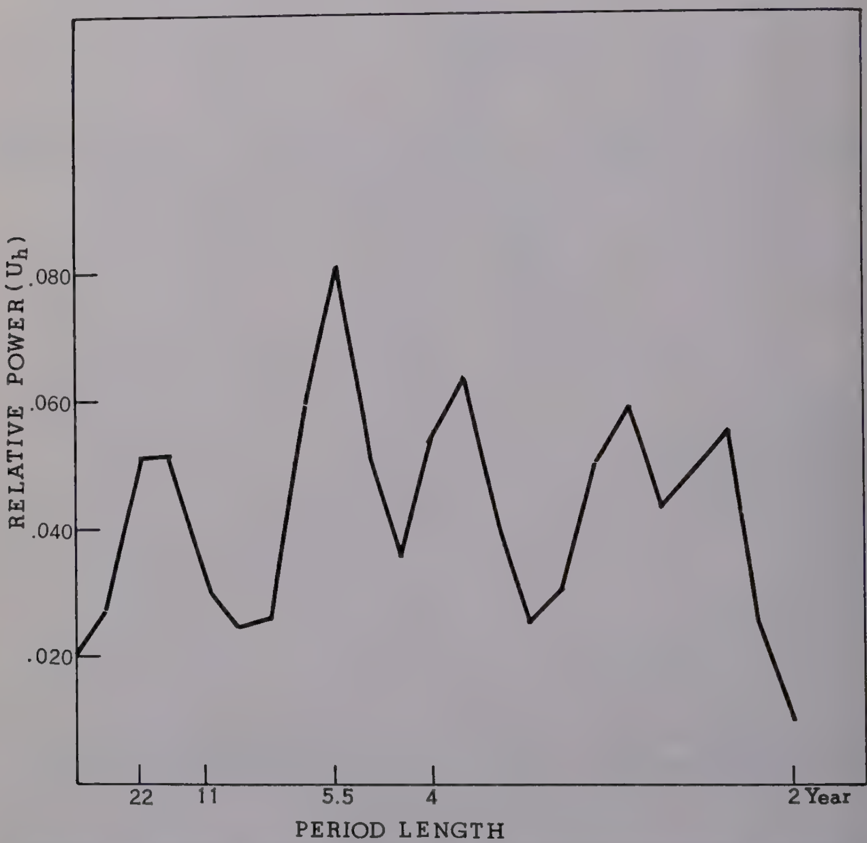


FIGURE 3. Spectral estimates (U_h) of the annual precipitation totals for Copenhagen, Denmark.

zero mean, the values are plotted here as deviations from 1000 mm., the approximate mean of the X_t series. FIGURE 7 shows a similar plot for Milan, Italy. In both of these plots the amplitude of the 11-yr. wave is rather small compared with the higher frequency fluctuations but, of course, this was anticipated on the basis of the power spectra results. In contrast, the application of the 11-yr. band-pass filter to the annual sunspot numbers (see FIGURE 8), shows relatively little reduction in amplitude of the original X_t series.

Plots similar to FIGURES 6 and 7 were constructed for each of the 6 other stations. The next step was to determine for each of the 8 stations the great-

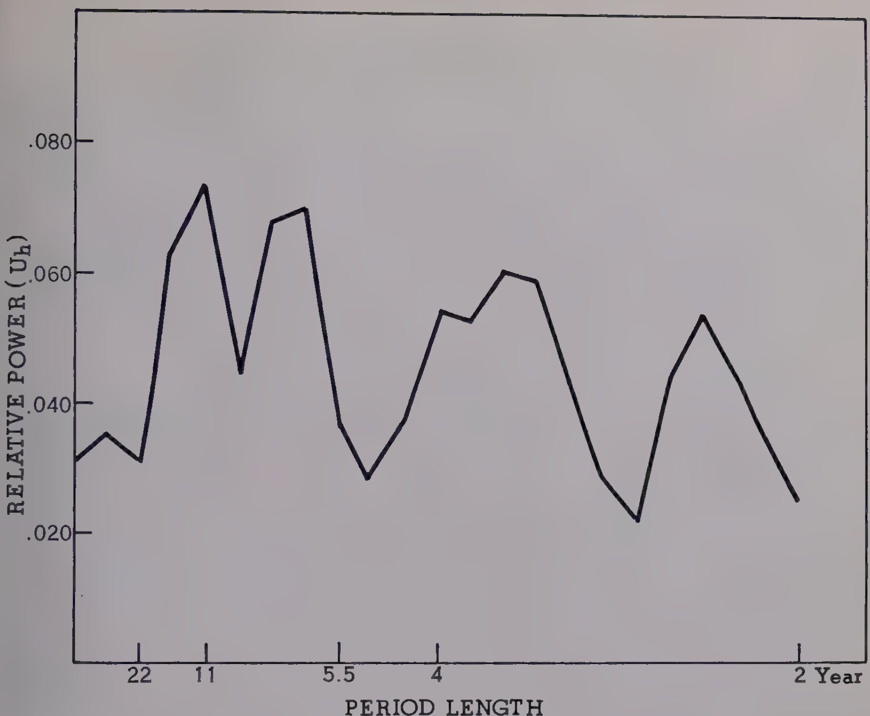


FIGURE 4. Spectral estimates (U_h) of the annual precipitation totals for Greenwich, England.

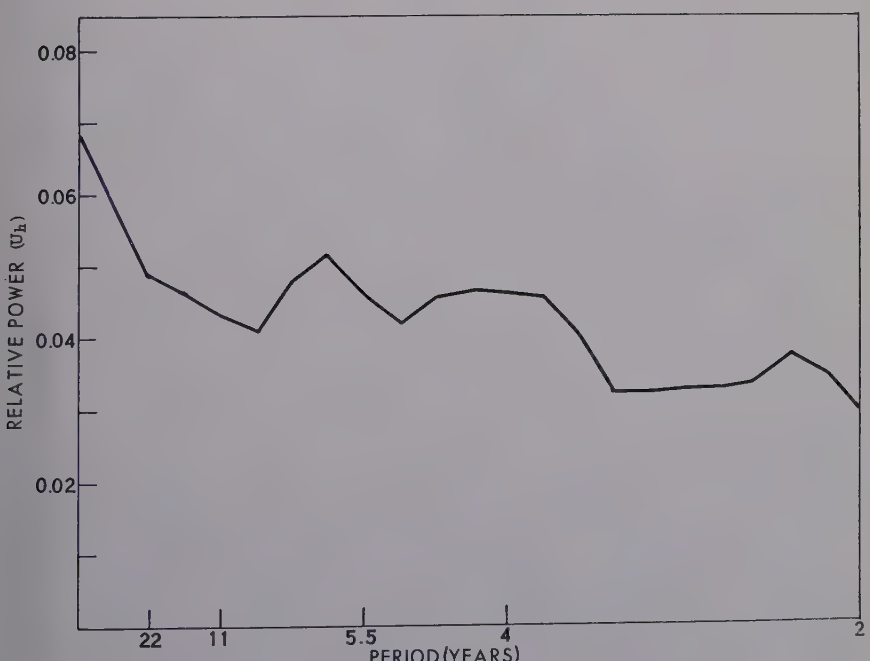


FIGURE 5. Spectral estimates (U_h) averaged for the 8 stations.

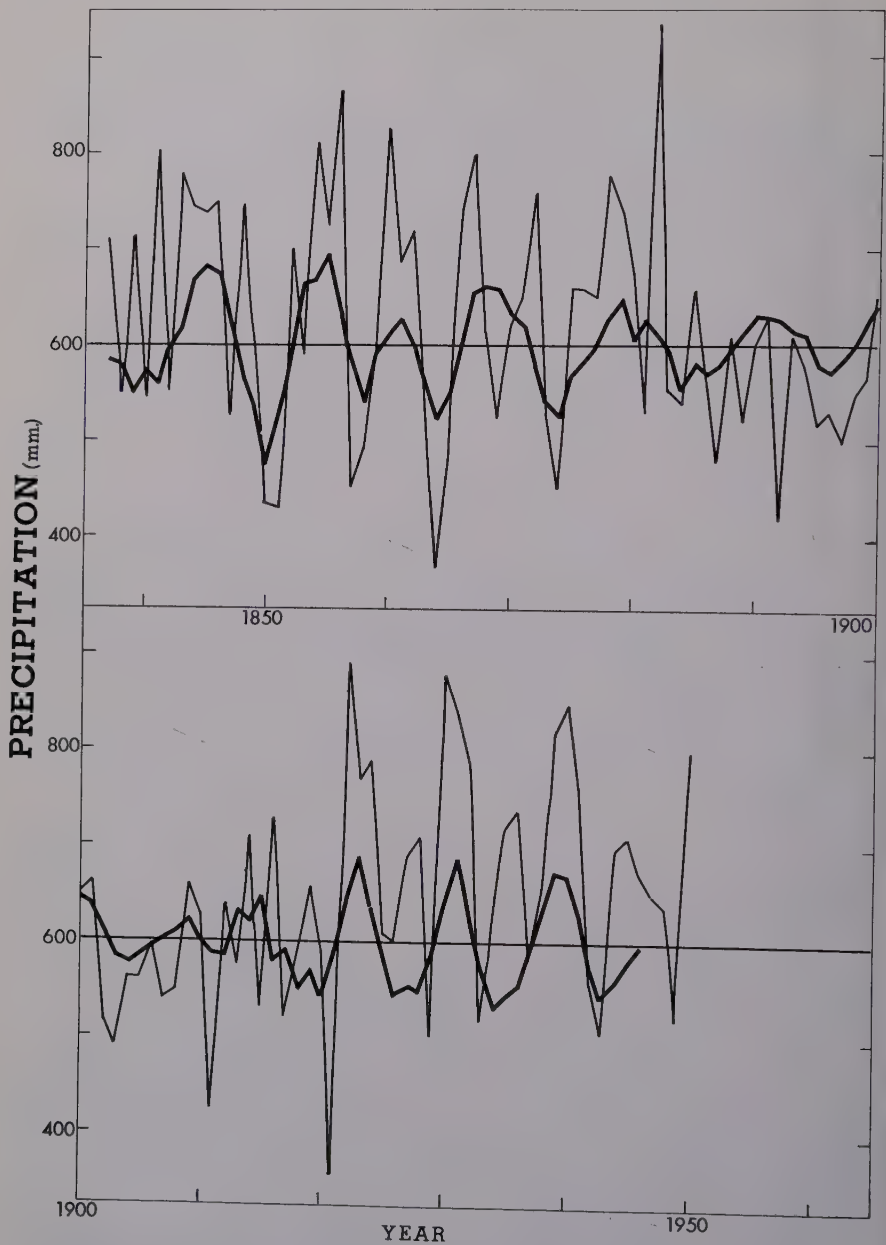


FIGURE 6. Annual precipitation totals for Frankfurt, West Germany (*thin line*) and the corresponding values of the filtered series (*heavy line*).

est positive anomaly (A) for the F_t series and identify the year in which it occurred. Thus for example in the Frankfurt record, the greatest positive anomaly was 94.7 mm. in 1855. This was called the highest peak where a peak was defined as a value higher than the 2 adjacent values. For each station the 10 highest "peaks" thus defined were determined and their dates and values tabulated. These dates were then converted into an index representing their locations relative to the sunspot cycle where $\Theta = 0°$ indicates the be-

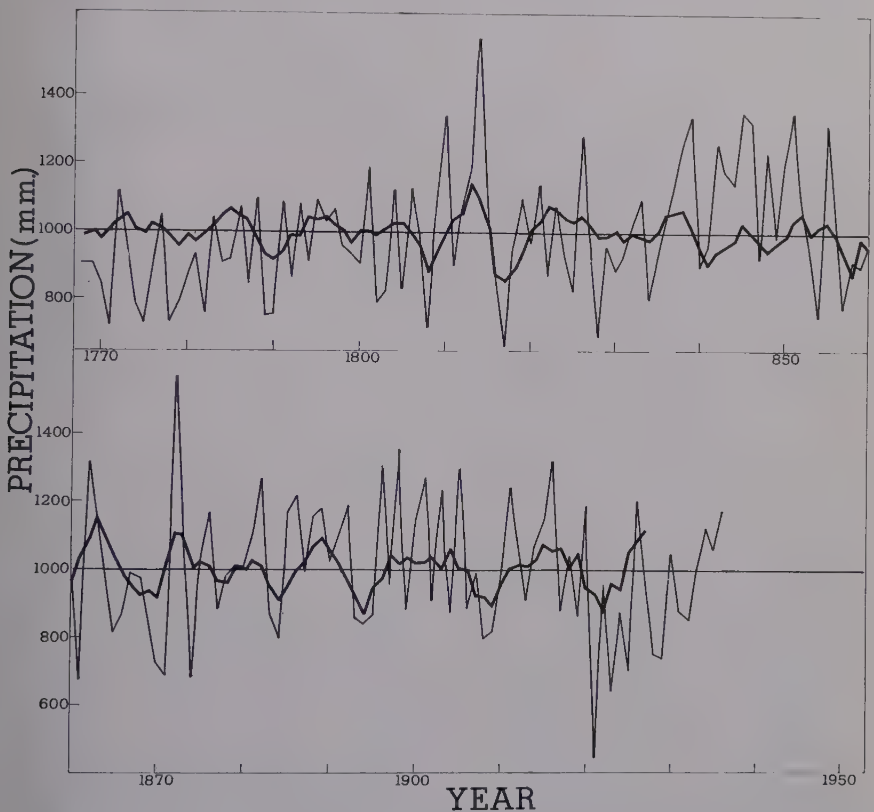


FIGURE 7. Annual precipitation totals for Milan, Italy (*thin line*) and the corresponding values of the filtered series (*heavy line*).

ginning of the cycle; $\Theta = 90°$ indicates 25 per cent of the way through the cycle; $\Theta = 180°$ indicates half way through the cycle, the formula continuing in this way. This was accomplished by constructing and using TABLE 3, which was based on the information given by Chernosky and Hagan.[1] The Zurich system considers a sunspot cycle as beginning at a minimum, and the cycles have been numbered consecutively 1, 2, 3, and up, beginning with 1755.

After the dates of the 10 highest peaks in each F_t series were expressed in terms of the sunspot cycle, the values of Θ and the anomaly A were plotted on a harmonic dial or sunspot clock shown in FIGURES 9 and 10 for Budapest,

Hungary, and Milan, Italy, respectively. The numbers plotted adjacent to the points refer to sunspot cycle numbers given in TABLE 3, in which c.g. means the center of gravity of the plotted points. The radii of the circles have been drawn arbitrarily at 100 mm. of precipitation to give some idea of the absolute amplitude of the fluctuations or anomalies. In these figures it is noticed that the sunspot maximum is shown somewhat before the cycle is half completed. This average condition is well known, of course, as is the fact that the length of the cycle and the relative position of the maximum varies somewhat.

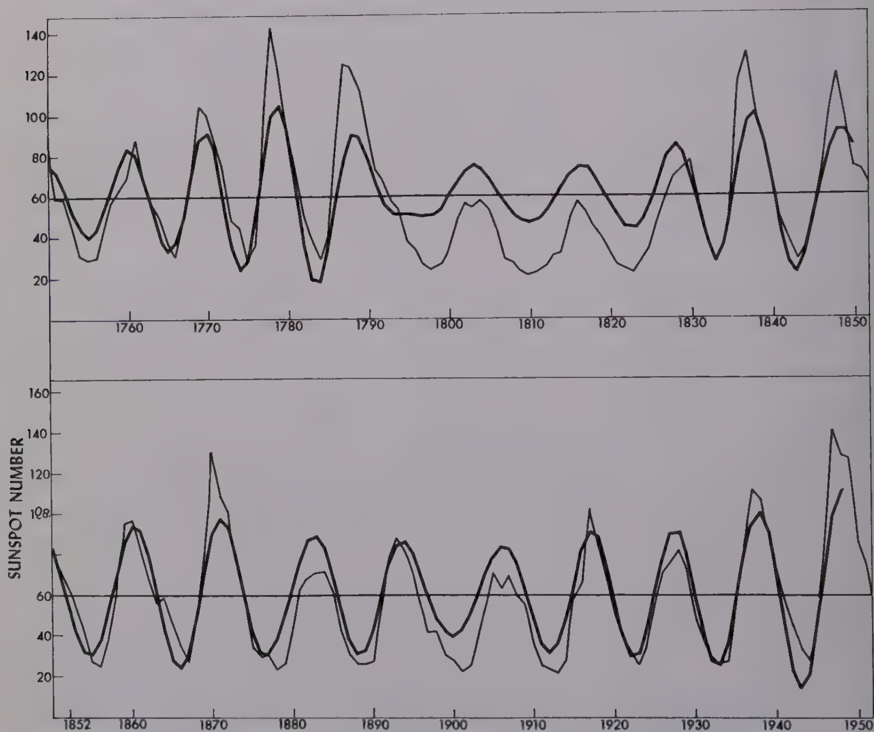


FIGURE 8. Annual Zurich sunspot numbers (*thin line*) and the corresponding values of the filtered series (*heavy line*).

Graphs similar to FIGURES 9 and 10 were prepared for the 6 other stations and the center of gravity determined. Tests are available to determine whether the center of gravity departs significantly from the center of the dial, and independence of the points can be assured by requiring that the plotted points refer to years separated by more than 20 yr. in time. Of the 8 stations studied, there was 1 highly significant ($P = 0.01$) departure for the center of gravity, and 1 other was on the borderline of significance. A summary of all 8 stations was prepared by plotting the phase Θ of the 2 highest peaks from each station on a harmonic dial. These results are shown in FIGURE 11. Since there appeared to be a slight tendency for the points to occur in the first half of the

TABLE 3

IDENTIFICATION OF YEARS ACCORDING TO ZURICH SUNSPOT CYCLE NUMBER AND RELATIVE POSITION (⊖) IN RESPECT TO THE BEGINNING OF THE CYCLE

Year	Zurich cycle No.	Degrees	Year	Zurich cycle No.	Degrees
1755	1	10	1810	5	357
1756	1	41	11	6	26
1757	1	73	12	6	54
1758	1	105	13	6	82
1759	1	137	14	6	111
1760	1	169	15	6	139
1761	1	201	16	6	167
1762	1	233	17	6	196
1763	1	264	18	6	224
1764	1	296	19	6	252
1765	1	328	1820	6	280
1766	2	0	21	6	309
1767	2	40	22	6	337
1768	2	80	23	7	7
1769	2	120	24	7	41
1770	2	160	25	7	75
1771	2	200	26	7	109
1772	2	240	27	7	143
1773	2	280	28	7	176
1774	2	320	29	7	211
1775	3	360	1830	7	244
1776	3	39	31	7	278
1777	3	78	32	7	313
1778	3	117	33	7	346
1779	3	157	34	8	23
1780	3	196	35	8	60
1781	3	235	36	8	98
1782	3	274	37	8	135
1783	3	313	38	8	172
1784	3	352	39	8	210
1785	4	21	1840	8	248
1786	4	48	41	8	285
1787	4	74	42	8	323
1788	4	101	43	9	360
1789	4	127	44	9	29
1790	4	154	45	9	58
1791	4	180	46	9	86
1792	4	207	47	9	115
7793	4	233	48	9	144
1794	4	260	49	9	173
1795	4	286	1850	9	202
1796	4	313	51	9	230
1797	4	339	52	9	259
1798	5	6	53	9	288
1799	5	35	54	9	317
1800	5	64	55	9	346
1801	5	94	56	10	16
1802	5	123	57	10	48
1803	5	152	58	10	80
1804	5	181	59	10	113
1805	5	211	1860	10	145
1806	5	240	61	10	177
1807	5	269	62	10	209
1808	5	298	63	10	241
1809	5	328	64	10	273

TABLE 3—*Continued*

Year	Zurich cycle No.	Degrees	Year	Zurich cycle No.	Degrees
1865	10	305	1908	14	206
1866	10	338	09	14	236
1867	11	9	1910	14	266
1868	11	40	11	14	297
1869	11	71	12	14	327
1870	11	102	13	14	357
1871	11	133	14	15	32
1872	11	163	15	15	68
1873	11	194	16	15	104
1874	11	225	17	15	140
1875	11	256	18	15	176
1876	11	286	19	15	212
1877	11	317	1920	15	248
1878	11	348	21	15	284
1879	12	20	22	15	320
1880	12	54	23	15	356
1881	12	88	24	16	32
1882	12	121	25	16	67
1883	12	155	26	16	102
1884	12	188	27	16	138
1885	12	222	28	16	173
1886	12	256	29	16	208
1887	12	289	1930	16	244
1888	12	323	31	16	279
1889	12	356	32	16	314
1890	13	27	33	16	350
1891	13	57	34	17	24
1892	13	86	35	17	59
1893	13	116	36	17	94
1894	13	146	37	17	128
1895	13	175	38	17	163
1896	13	205	39	17	197
1897	13	235	1940	17	232
1898	13	265	41	17	267
1899	13	295	42	17	301
1900	13	324	43	17	336
1901	13	354	44	18	107
1902	14	24	45	18	46
1903	14	55	46	18	82
1904	14	85	47	18	118
1905	14	115	48	18	153
1906	14	145	49	18	189
1907	14	176	50	18	225

solar cycle, another plot was made for the third and fourth highest peaks, as shown in FIGURE 12. In this case, 14 of the 16 points occurred in the first half of the solar cycle. A number of interesting coincidences are suggested by an examination of these 2 graphs.

Conclusions

Although it may be premature to draw any general conclusions, the results of the power spectra analyses indicate that no appreciable component of the 11-yr. solar cycle shows up in the precipitation data. The objections may be made of course that only annual precipitation totals have been used, and that it would be much more proper to use seasonal values since any possible solar

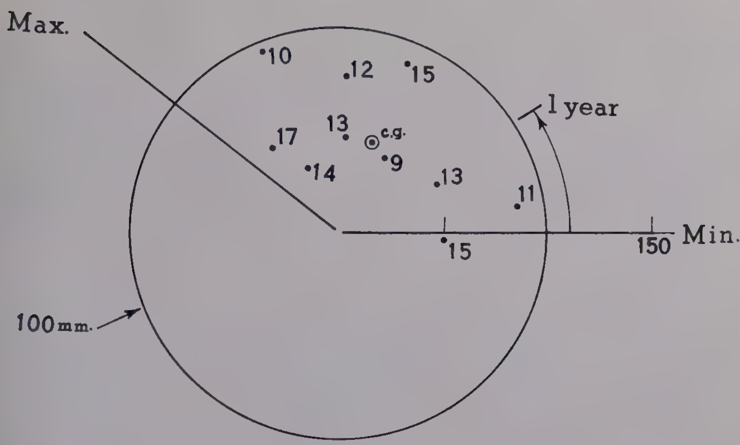


FIGURE 9. Harmonic dial showing phase and amplitude of the greatest positive anomalies in the filtered series derived from the annual precipitation totals (Budapest, Hungary). The numbers are the Zurich cycle numbers.

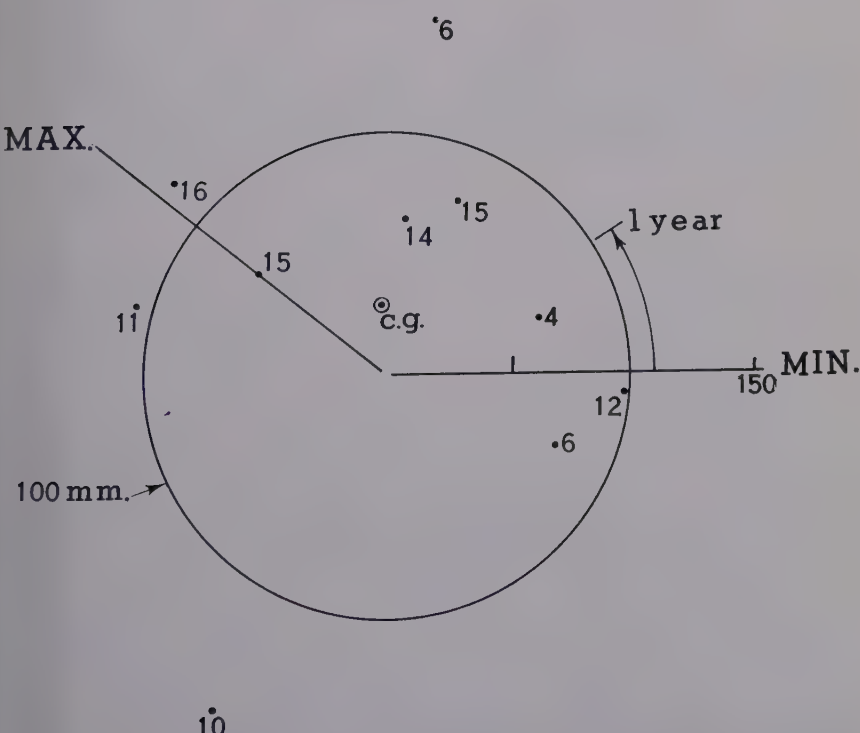


FIGURE 10. Harmonic dial showing phase and amplitude of the greatest positive anomalies in the filtered series derived from the annual precipitation totals (Milan, Italy). The numbers are the Zurich cycle numbers.

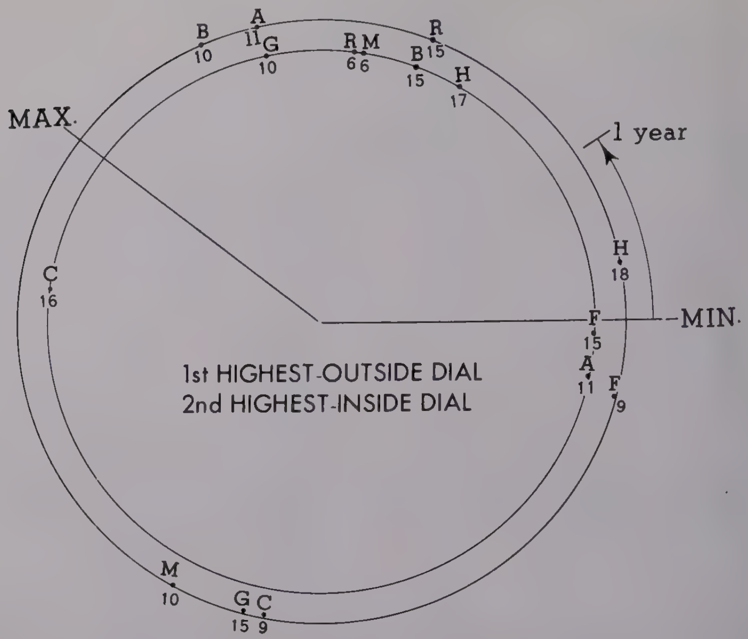


FIGURE 11. Summary dial showing position in sunspot cycle of highest and second-highest anomalies in filtered series for the 8 stations. Key for stations: (*A*) Albany, N.Y.; (*B*) Budapest, Hungary; (*C*) Copenhagen, Denmark; (*F*) Frankfurt, West Germany; (*G*) Greenwich, England; (*H*) Helsinki, Finland; (*M*) Milan, Italy; and (*R*) Rome, Italy.

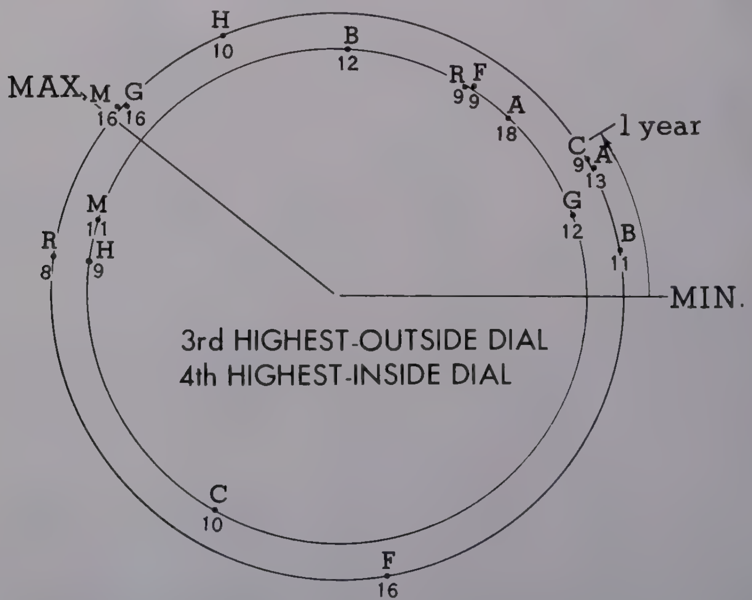


FIGURE 12. Summary dial showing position in sunspot cycle of third- and fourth-highest anomalies in filtered series for the 8 stations.

control would operate differently according to the time of the year. There is the possibility that the solar effect does not always act in the same direction but depends upon the initial conditions of the earth-atmosphere system in some nonlinear fashion. That some such interaction may exist is suggested by the results of examining the anomalies in the filtered precipitation series where some evidence of nonrandomness in phases is indicated. However, this is not clear, and an attempt at further interpretation or speculation should wait for the completion of the statistical analysis.

References

1. Chernosky, E. J. & M. P. Hagan. 1958. The Zurich sunspot number and its variations for 1700–1957. J. Geophys. Research. **63:** 775–788.
2. Tukey, J. W. 1949. The sampling theory of power spectrum estimates. Symposium on Applications of Autocorrelation Analysis to Physical Problems. Woods Hole, Mass. : 47–68.
3. Wiener, N. 1930. Generalized harmonic analysis. Acta Math. **55:** 117–258.
4. Wiener, N. 1949. Extrapolation, Interpolation, and Smoothing of Stationary Time Series. Technology Press of M.I.T. Cambridge, Mass.
5. Ward, F. W., Jr. 1960. The variance (power) spectra of C_i, K_p, and A_p. J. Geophys. Research. **65:** 2359–2373.
6. Panofsky, H. A. & G. W. Brier. 1958. Some Application of Statistics to Meteorology. Pa. State Univ. University Park, Pa.
7. Blackman, R. B. & J. W. Tukey. 1958. The measurement of power spectra from the point of view of communications engineering. Bell System Tech. J. **37:** 185–282, 485–569.
8. Holloway, J. L., Jr. 1958. Smoothing and filtering of time series and space fields. Advances in Geophysics. **IV:** 351–389. Academic Press. New York, N.Y.
9. Clayton, H. H. & F. L. Clayton. 1944, 1947. World Weather Records. Smithsonian Misc. Collections, 79, 90, 105. Smithsonian Institution. Washington, D.C.
10. United States Weather Bureau. 1959. World Weather Records, 1941–50. U. S. Gov't. Printing Office. Washington, D. C.

Part III. Meteorology and Climatology

THE GENERAL CIRCULATION OF THE ATMOSPHERE AS A NECESSARY LINK IN THE SUN-CLIMATIC VARIATIONS CHAIN

B. L. Dzerdzeevskii

Institute of Geography, U.S.S.R. Academy of Sciences, Moscow, Union of Soviet Socialist Republics

Introduction

This paper treats some of the problems connected with climatic variations in the first half of the 20th century. Consequently it deals with only a small part of the general problem of climatic variations, the complexity and broadness of which is demonstrated by the contents of this publication.

Nevertheless it seems worthwhile to analyze the processes involved even for such a short period of time. It is from the beginning of this century that synoptic weather maps have been available, making possible a more detailed analysis of the phenomena in question and paving the way for working out a theory for forecasting climatic variations over periods of two or three decades.

It is generally accepted that short-period climatic variations are related to the variable effect of solar activity. This relationship is investigated along two lines: (1) elucidation and study of the transfer mechanism of solar effects to the earth; (2) elucidation and study of the processes of the earth's atmosphere dependent on these effects and the influence of such processes on climate.

It is generally recognized that the main difficulties are inherent in the first task. This tacitly implies that in the second, purely atmospheric, task the situation is better understood, but this is not the case.

The chain of relationships to be established—sun → climate-forming processes → climate → weather—should be continuous. Consequently atmospheric processes should be represented in the chain only once. This means that in analyzing both the way in which solar effects are received by the earth as well as the direct "earthly" causes of climatic variations induced by these effects the same atmospheric processes should be considered. The general circulation of the atmosphere adequately meets the requirements of such an intermediate link.

It should be emphasized that by circulation is meant the *general* circulation of the earth, that is, the planetary circulation, or the circulation over at least a hemisphere.

It is common knowledge that not only the links among weather, climate, and general circulation, but also those between weather and individual synoptic processes (patterns) are not fixed but change in time. This variability constitutes the chief source of the difficulties inherent in the second, atmospheric part of the climatic variation problem.

This paper reports some of the results of the work aimed at overcoming these difficulties.

Discussion

In discussing the connection of weather, climate, and atmospheric processes it is worthwhile to recall some of the results of investigation that confirm the *similitude* of the links between the weather and individual synoptic patterns with the links between climate and the general circulation. Both components of the last pair should be treated dynamically just as those of the first pair should be. Hence what we have in mind here is the analysis of simultaneity in the changes of circulation characteristics and the variations of climatic elements.

The extensive material that has accumulated testifies to the fact that the atmospheric processes of a lower order are to be considered as fluctuations with respect to the processes of a higher order, that is, the diurnal changes are fluctuations with respect to seasonal and annual changes, while seasonal and annual changes are fluctuations with respect to those found in long-period data. The relationship between these categories of processes is determined from stochastic ties.

The material for studying the dynamics of the large-scale circulation processes at the atmosphere is usually prepared by two methods: (1) typing of the synoptic weather patterns; (2) calculation of the circulation indices. It is more difficult than ordinary typing to obtain a numerical expression for the types of large-scale, atmospheric patterns, but this method yields a more graphic characterization of the general and particular links between the atmospheric patterns and better reflects their relation to the conditions of the earth's surface. If the typing is rational it is easily brought into agreement with the circulation indices.

Any typing of atmospheric patterns inevitably entails generalization of the features to be typed. The lowest limit of such generalization is a well-pronounced seasonal change of the types. Only however in this case can a required dynamic analysis of the patterns be secured. A generalized annual scheme does not provide possibilities for such analysis.

The most important and, at the same time, the most convenient basis for typing the atmosphere circulation patterns over a hemisphere is provided by the ratio between the zonal and meridional components of the circulation.* Since the material required for statistical analysis is available only in the form of surface synoptic maps of the northern hemisphere, and since the fundamental steering currents in the middle troposphere are satisfactorily described by the tracks of most cyclones and anticyclones it is convenient to use these tracks as indices of zonality and meridionality in the patterns over the hemisphere.† In the northern hemisphere (as a matter of fact, in its extratropical latitudes, since in the tropical zone the number of observation points is considerably smaller) a particularly important role is played by intrusions (flow

* Meridional components are directed northward or southward; zonal circulation from west to east.—Ed.

† Thus a zonal circulation process or pattern would be marked by eastward motion of the surface cyclones and anticyclones. North-south orientation of the tracks of the cyclones and anticyclones would indicate a meridional circulation pattern.—Ed.

from the Arctic region) responsible for blocking actions in the west-east zonal circulation.

Although the duration of a single macroprocess (large-scale circulation pattern) as calculated by different investigators, varies in the majority of cases between four or five days, this figure should not be used as a constant in a further, more detailed analysis. One should seek to establish actual, natural limits for each individual process. Apart from making the analysis more detailed, it also permits the calculation of results pertaining to the duration of the various processes in analyzing the dynamics of their development, which is of particular value in studying climatic variations. A transition from one circulation type over the hemisphere to another occurs very rapidly, so that synoptic maps of two borderline days may, as a rule, be referred to different circulation types.

On the basis of all the above-mentioned principles my co-workers and I have developed a typing of the atmospheric circulation over the northern hemisphere for 56 years of the 20th century (using more than 20,000 synoptic maps). It was found possible to reduce the great variety of synoptic processes to 13 principal circulation types.

Principles of the typing, comparison of the variation of circulation types and circulation indices, determination of the borders and structure of natural synoptic seasons, as well as some of the results obtained in analyzing variability of circulation and climatic indices are available in greater detail elsewhere.[1-13]

All this material is interesting in view of the fact that the covered period closely coincides with the ascending branch of the current secular cycle of solar activity. It makes possible the comparisons and analyses required by the program of the first task related to the problem in hand (investigation of the way in which solar effects are received by the earth).

Extensive and effective research along these lines has been carried out by A. Bezrukova and B. Rubashev at the Astronomical Observatory in Pulkovo, Union of Soviet Socialist Republics. In a number of papers these investigators have reported results obtained through a detailed comparison of the above-mentioned circulation types over the northern hemisphere with the long-period variations of solar activity (the ascending branch of the secular cycle), with its 11-year cycle, as well as with geomagnetically perturbed and geomagnetically calm days.

As an index of solar activity Bezrukova and Rubashev have chosen a relative number of sun-spots and their total area recently calculated by them separately for the northern and southern hemispheres of the sun. This choice is determined by the fact that long-series observations have been performed for this index and are not available for other indices. At the same time the connection between various solar indices is rather close.

Investigating the dependence of all the 13 circulation types upon the mentioned indices of solar activity by Bezrukova[2] established a good agreement among them. Direct correlation is observed for all the types of zonal motion and weak violations of zonal flow; inverse correlation is inherent to the types of meridional circulation.

According to this main pattern all the circulation types differing in minor details can be united into two groups: zonal and meridional. FIGURE 1, re-

produced from Bezrukova's paper, shows the corresponding curves. The secular amplitude of the zonal circulation types (together with types showing violation of zonality) increased 2.2 times, while that of meridional ones dropped by the factor of 1.6; the amplitude of the solar activity index for the same period increased twofold.

Investigating the connection of the zonal circulation in winter with the 11-year cycle of solar activity Bezrukova[13] has also discovered good agreement

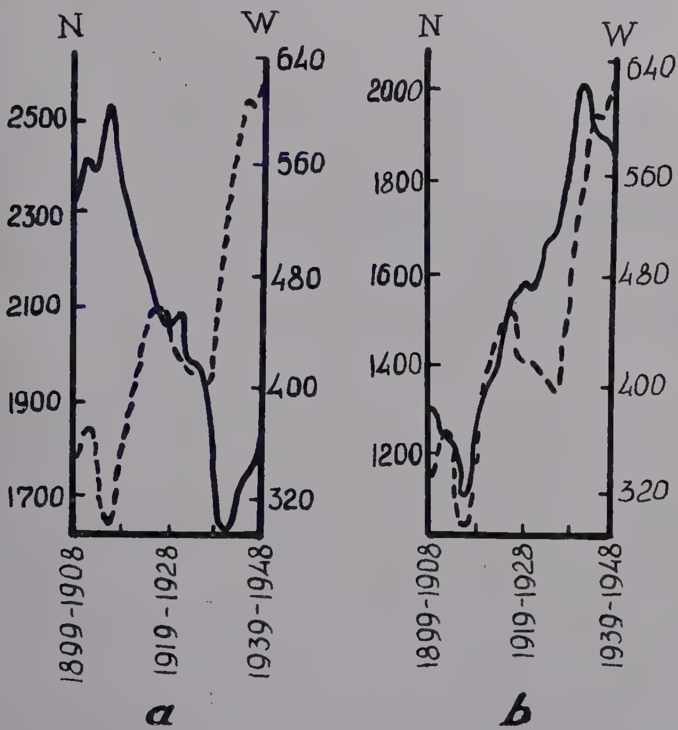


FIGURE 1. Long-period changes of solar activity and variations of atmospheric circulation. (*a*) Fluctuations in frequency of the meridional types of circulation. (*b*) Fluctuations in frequency of the zonal types of circulation. (*n*) Duration (in days) of Dzerdzeevskii's circulation types. (*w*) Number of sunspots. Circulation of the atmosphere (——); solar activity (– – – –).

between the two curves (FIGURE 2). It is noteworthy that during cycle 16 and 18 the curves have two peaks. Peaks in the zonal circulation curve were observed in one year after the maximum total area of sun-spots was reached, and in two years after the maxima of the two-peak curves of the sunspots area.

Rubashev in a number of his works[8,10,12] has investigated the short-period links of the same 13 general circulation types with solar activity including cases of geomagnetically active and geomagnetically calm days. Although these problems are directly related to weather conditions rather than climate it is worthwhile to mention them, since Rubashev's work helps to define more specifically the relationships between the sun and the earth's atmosphere.

In 1960 Rubashev[12] showed that a delayed occurrence of a circulation "anomaly" (that is, the case when average values for a given circulation group exceed the standard deivation) with respect to the solar extremes ranges from 3 to 6 months, and that positive "anomalies" of the meridional circulation types (including cases of violation of zonality) are connected with the minima of solar activity.

Investigating relationships between the circulation types and the geomagnetically perturbed and calm days Rubashev[8] has established that the zonal

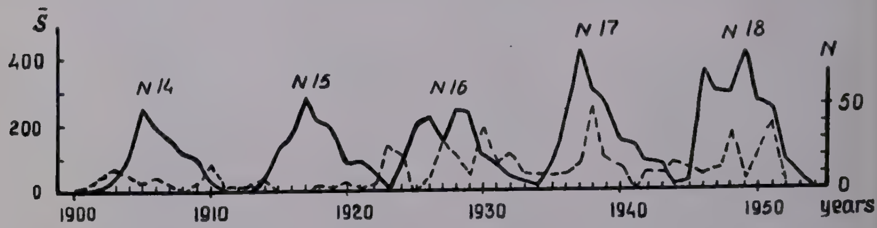


FIGURE 2. Annual sums of the total average areas of sunspots in the sun's northern hemisphere (— —) and annual total number of days with zonal circulation in the earth's atmosphere in winter (- - - -).

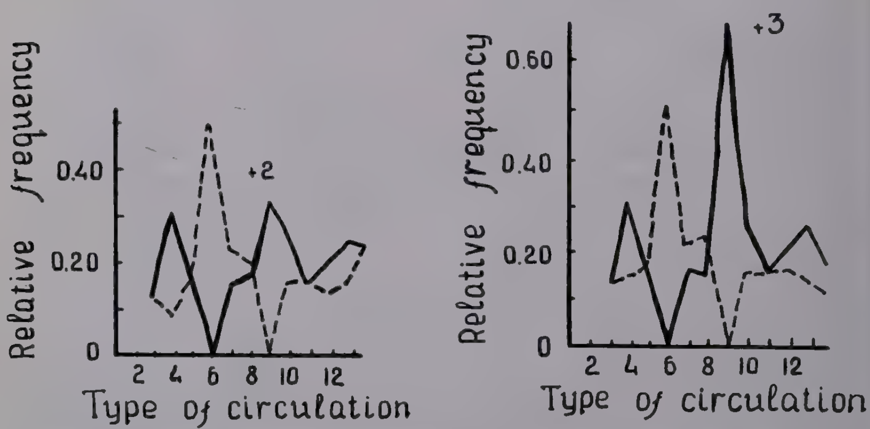


FIGURE 3. Curves of the relative frequency of the circulation types with respect to geomagnetically perturbed (——) and geomagnetically calm (- - - -) days.

circulation types and arctic intrusions into the Pacific Ocean are usually observed near magnetically calm days, while violations of zonality due to Arctic intrusions into the Atlantic and Europe occur near magnetically perturbed days (FIGURE 3).

Unfortunately only this brief mention can be made here of these interesting and important investigations, which illustrate the connection between our general circulation types and solar activity variations.

We shall now pass to a comparison of *the same* circulation characteristics with climatic indices in individual points of the hemisphere.

It has already been observed that we have performed all possible calcula-

tions of circulation characteristics in the northern hemisphere according to our system of types including such factors as frequency and duration of zonal and meridional circulations, connection with circulation indices, and deviations from long-period means (assumed to be equal to zero). All the calculations were made for separate months, seasons and years, as well as for 5-, 10-, and 20-year running means. We were also compelled to increase the number of "synoptic seasons" of the year up to 6 (prespring, spring, summer, autumn, prewinter, winter).

All the calculations were made for the circulation characteristics of the entire hemisphere as a whole. Thus the presence of a single Arctic intrusion irrespective of its direction (for example, into the Atlantic Ocean, into Europe, or into the Pacific Ocean) was taken to indicate violation of zonality. If the number of intrusions ranged from 2 to 4, also irrespective of their geographic position, this indicated meridionality.

Simultaneously all the calculations were repeated for the seven sectors into which the Northern Hemisphere was divided. This made it possible to derive more specific circulation characteristics for individual areas. It should be taken into account that a single intrusion over the hemisphere was regarded only as a violation of zonality, since zonal motion is preserved over the predominant part of the hemisphere. However in the area where the intrusion occurred we are faced with meridional tracks.

It should be emphasized that in all cases the initial material was provided by the same synoptic maps of the *northern hemisphere*; a complete set of these was included in the calculations eight times. This work was performed on computors.

In addition to circulation characteristics, temperature, pressure, and precipitation calculations were performed in a similar way (as indicated above) for all the years of observation at all the stations listed in *World Weather Records*.[14]

In this way uniformity of the material was achieved that made possible the carrying out of necessary comparisons of the variations of the climatic-elements variations at individual points with the circulation characteristics of the whole hemisphere. For this purpose we first plotted graphs reflecting all circulation characteristics and the long-period variations of climatic indices for all the observation points. The graphs were plotted on the basis of the direct data, 5-, 10-, and 20-year overlapping sums and means; integral curves were also constructed.

Direct (or reverse) dependence of long-period variations of climatic indices at individual points of the hemisphere upon circulation characteristics of the whole hemisphere is found to be sufficiently good everywhere. It is better expressed in high latitudes and with respect to temperatures and is somewhat less prominent as regards precipitation. This result is to be attributed to a greater diversity of circulation processes in middle latitudes and to the fact that air-mass precipitation, which often points to a strongly developed anticyclonic regime, seemingly violates the agreement between the curves of circulation and climatic characteristics although, in fact, they only emphasize the connection between them. A similar result, although not so well expressed, happens to temperature during winter radiational cooling in anticyclones.

The correlation coefficients of the analyzed dependence vary from 0.59 to 0.92, the prevailing value being above 0.75 ± 0.03 to 0.04.

Figures illustrating the comparisons carried out for a number of observation points have been published in previous papers of mine.[5,9] FIGURES 4 to 6 of this paper contain several analogous illustrations. The principles of their plotting were expounded above, and the captions provide necessary explanations. Therefore there is no need to give additional description and analysis here. I shall only make the following general remarks.

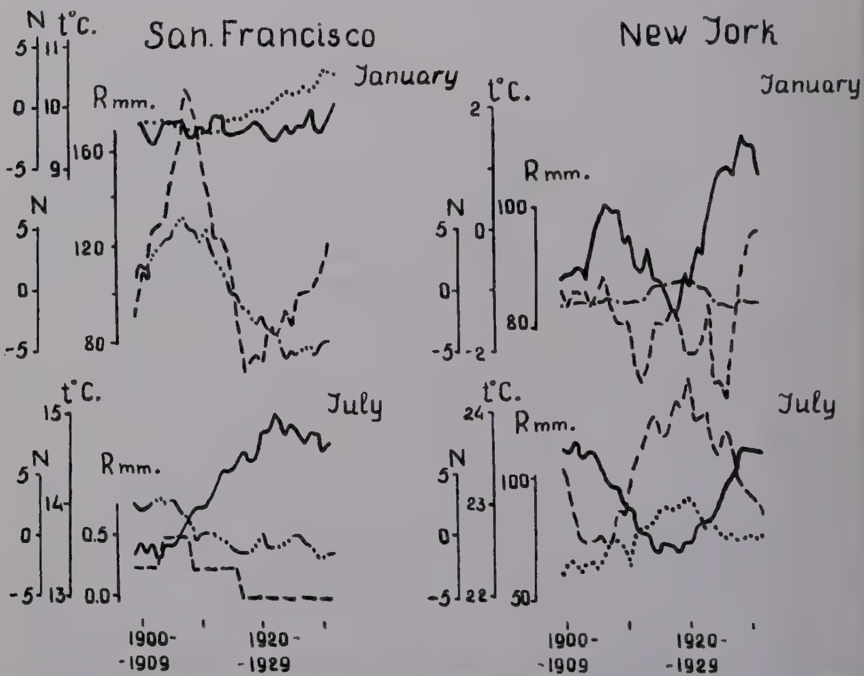


FIGURE 4. Temperature and precipitation curves at San Francisco, Calif., and New York, N. Y., stations, and circulation types of the northern hemisphere in January and July: temperature (——), precipitation (– – – –), zonal circulation (· · · · · ·), violation of zonality (—· —), meridional circulation (—· · · —).

It is evident that intensification or weakening of the zonal circulation does not produce identical changes of temperature and precipitation on the western and eastern coasts of the continents, and that the meridional circulation effect is different in the Atlantic, in Siberia, and elsewhere. In many cases it is possible to single out the type of circulation that at a given place and for *a given season* proves to be the prevailing or, even, the chief one (see, for instance, Kazalinsk station in FIGURE 6). To simplify the figures it is only such a "dominant" curve that is plotted on the majority of them. However this is only for the sake of illustration. In our analyses we have certainly used all the components of the circulation.

Numerous researchers have already pointed out the change in the character

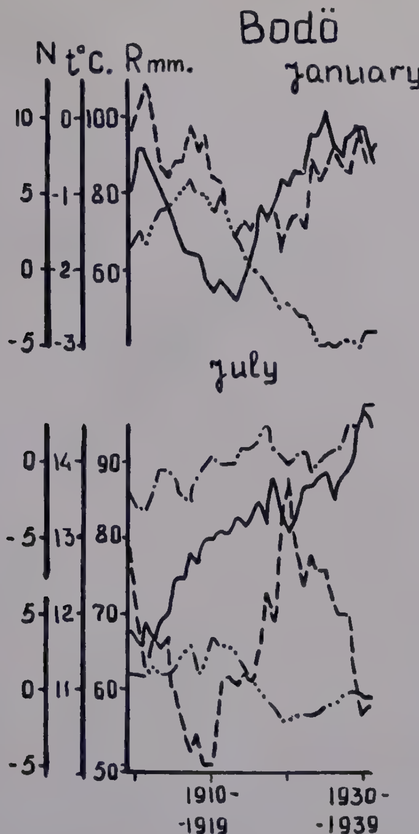


FIGURE 5. Temperature and precipitation curves at Bodö station, Norway, and circulation types of the northern hemisphere in January and June. Temperature (——), precipitation (– – – –), violation of zonality (—· —), meridional circulation (—· · · —).

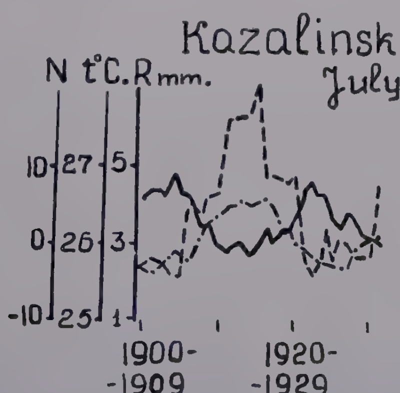


FIGURE 6. Temperature (——) and precipitation curves (– – · – –) at Kazalinsk station, in Soviet Union, in July, and violation of zonality in the northern hemisphere (– – · —).

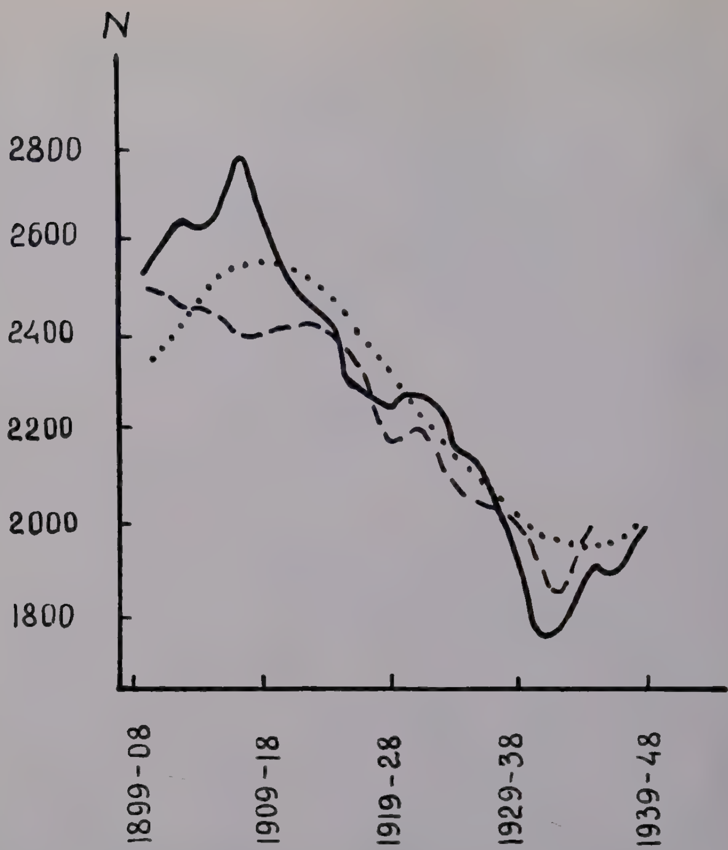


FIGURE 7. Changes in the levels of the lakes Michigan (– – – –) and Chany (· · · · · ·), and the march of meridional circulation type over the northern hemisphere (——).

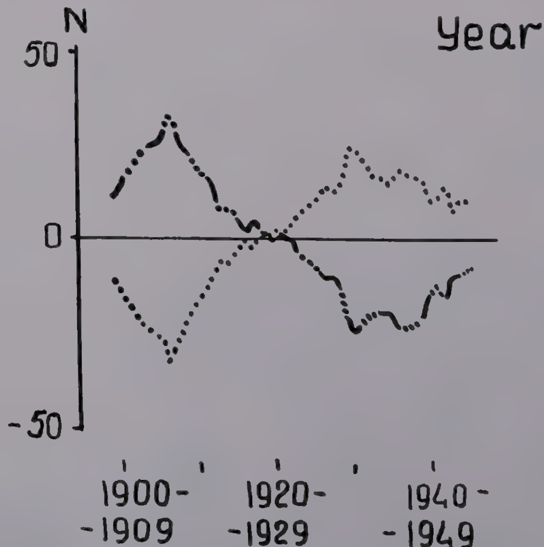


FIGURE 8. Long-period march of the zonal (· · · · · ·) and meridional (—· · · —) components of the atmospheric circulation in the northern hemisphere (annual data).

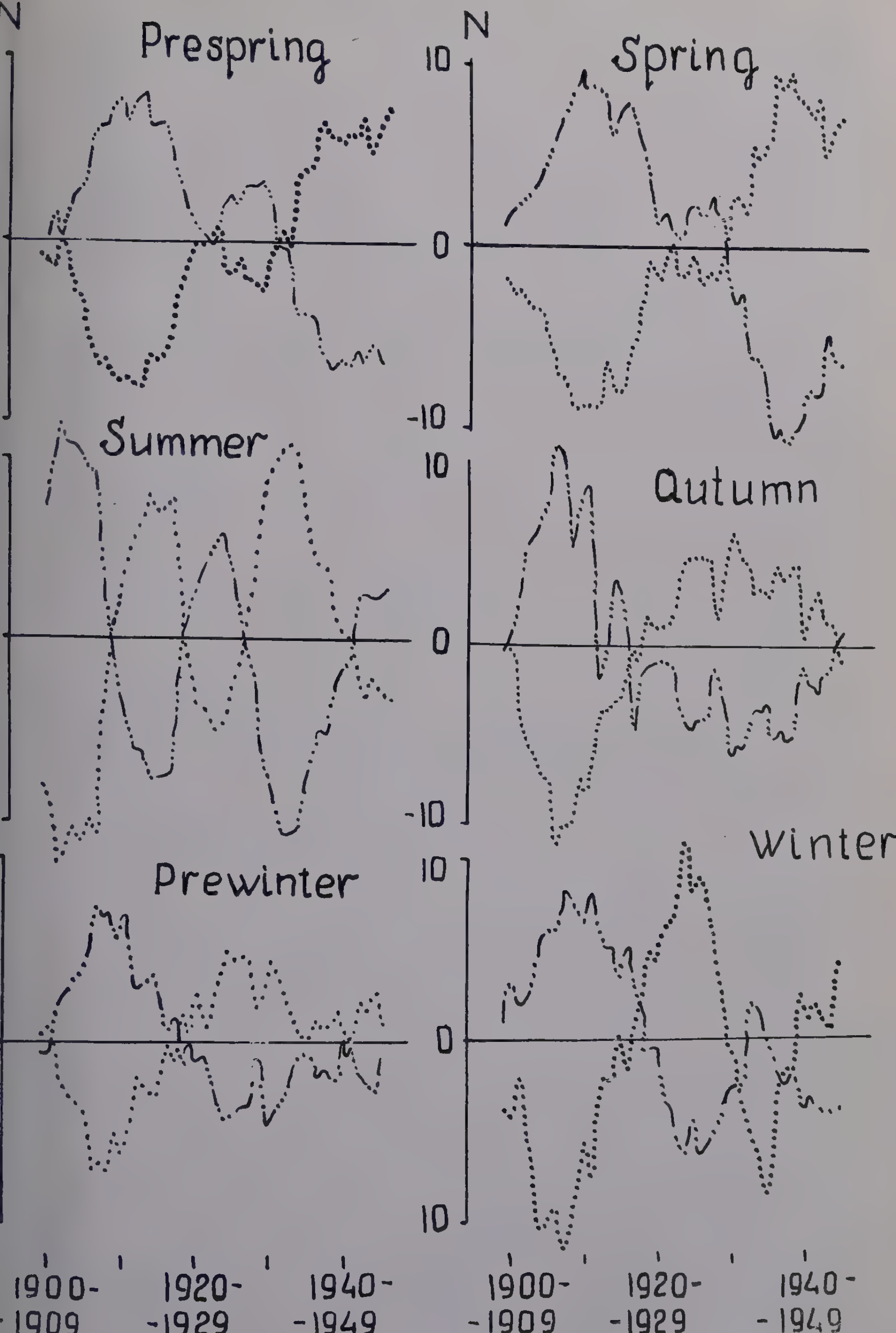


FIGURE 9. Long-period march of the zonal (· · · · · ·) and meridional (—· · · —) components of the atmospheric circulation in the northern hemisphere (6-season data).

and type of the dependence between variations in the climatic elements and circulation characteristics or solar activity indices observed thus far approximately since the 1920s. Our data in many cases confirm this. From the graphs plotted here, the Bodö station, Norway, may serve as an example (FIGURE 5, January).

The validity of the short-period climatic fluctuations considered above is confirmed by an extensive literature. To avoid accidental errors occasioned by the technique involved in processing the material, the calculation of the entire data was performed for periods of various duration. This validity is also logically confirmed by the above-mentioned local links of the circulation to the underlying surface conditions.

Bezrukova[4] has investigated fluctuations in the levels of big lakes (Michigan, Chany, Victoria, the Caspian Sea, and those in the Soviet Union). In all cases she obtained a fair agreement between lake levels and changes in frequency and duration of the circulation types. FIGURE 7 reproduces one of her graphs illustrating a close connection of the fluctuations in lake levels (Michigan and Chany) with a long-period march of meridional circulation types. Both lakes are situated at a tremendous distance from one another, and the simultaneity of their fluctuations can be attributed only to a common cause, that is, the general circulation of the atmosphere.

It is my opinion that the foregoing permits the hope that the circulation characteristics of the northern hemisphere developed by my co-workers and me may serve as a missing but necessary link in the sun-climate fluctuations chain.

It is widely known that there was a well-pronounced change in the character of the atmospheric circulation that occurred approximately in the 1920s. It is connected with a drastic transition from the prevalence of meridional circulation patterns to that of zonal processes (FIGURES 8 and 9). Lack of reliable data in the 19th century does not allow us to establish the beginning of this circulational and climatic epoch with sufficient accuracy. On the basis of some indirect data it is possible to date it back to the 1890s. Hence the duration of this epoch is close to 20 to 30 years.

A reverse change in circulation characteristics is observed in the 1940s. However, it is not so clearly and definitely expressed.

Analysis of the causes responsible for this change is beyond the limits of the present report.

References

1. DZERDZEEVSKII, B., V. KURGANSKAIA & Z. VITVITSKAIA. 1946. Tipizatsiia Tsirkulatsionykh Mekhanizmov v Servernom Polusharii i Kharakteristica Sinoptitcheskikh Sezonov. Gidnometeorologitcheskoie Izdatelstvo, Moskva. (Typing of the Circulation Mechanisms in the Northern Hemisphere and Characteristics of Synoptic Seasons. Moscow, U.S.S.R.)
2. BEZRUKOVA, A. 1950. Kharakter tsirkulatsii zemnoi atmosphery i solnetchnaja activnost. Bull. Komissii po issledovaniju solntsa, **5–6** (19–20): 75–84. (Characteristics of circulation of the Earth's atmosphere and solar activity. Bull. Commission of Solar Investigation. **5-6** (19–20): 75–84.)
3. DZERDZEEVSKII, B. & A. MONIN. 1954. Tipovyie skhemy obstchei tsirkulatsii atmosphery i index tsirkulatsii. Izvestia Akad. Nauk SSSR, Ser. geophys. **6:** 562–574. (Typical schemes of general circulation of the atmosphere and circulation indices. Proc. Acad. Sci. USSR, Geophys. Ser.)

4. Bezrukova, A. 1954. Vlianie Solnetchnoi Activnosti i Kharaktera Atmosphernoi Tsirkulatsii na Kolebania Urovnia Ozior i na Zasukhi. Trudy Laborat. Ozerovedenia Akad. Nauk SSSR.III.: 23–46. (The influence of solar activity and of characteristics of atmospheric circulation upon level of lakes and droughts. Ann. Labor. Lakes' Invest.)
5. Dzerdzeevskii, B. 1956. Problema Kolebanii Obstchei Tsirkulatsii Atmosphery i Klimata. In the books "Voeikov i Problemy Sovremen. Klimatologii, Gidrometeor. Izdat. Leningrad. **1956:** 109–122. (Problems of Variations of General Circulation of the Atmosphere and Climate.)
6. Dzerdzeevskii, B. 1956. Role d'analyse de la circulation atmospherique général pour l'établissement des frontiéres entre régions arides et humides. In Essais de Géographie, Edit de l'Acad. d. Sc. de l'USSR, Moscou, **1956:** 150–156.
7. Dzerdzeevskii, B. 1957. Tsirkulatsifonnye skhemy sezonov goda. Izvestia Akad. N. SSSR. Ser. Geograph. **1:** 36–55. (Schemes of seasonal circulation.)
8. Rubashev, B. 1958. O sopastavlenii reaktsii atmosphernoi tsirkulatsii i baritcheskikh polei na kolebanniia geomagnitnoi aktivnosti. Bull. "Solnetchnyie Dannyie" 1957 goda. **6:** 117–120. (On the comparison of reaction of atmospheric circulation and atmospheric pressure fields to fluctuations of geomagnetic activity.)
9. Dzerdzeevskii, B. 1959. Problemy Klimatologii Arktiki. In the book "Problemy Severa." Akad. Nauk SSSR. **3:** 168-179. (Problems of the arctic climatology.)
10. Rubashev, B. 1959. O raspredelenii otnositelnykh chastostei tsirkulatsionnykh mekhanizmov B.L. Dzerdzeevskogo vblizi geomagnitno-vozmustchennykh i geomagnitno-spokoinykh dnei. Bull. "Solnetch. dann." 1959 goda. **4:** 80–81. (On the distribution of relative frequencies of Dzerdzeevskii's circulation mechanisms near geomagnetically perturbed and geomagnetically calm days.)
11. Dzerdzeevskii, B. 1960. Half-age vaiiability of climate at high latitudes of the northern hemisphere and some problems of climatological classifications. XIX Intern. Geograph. Congr. 1960. Stockholm, Sweden.
12. Rubashev, B. 1960. Vnutrigodovyie fluctuaftsi solnechnoi activnosti i otriftsatelnye anomalii atmosfernoi tsirkulatsii. Bull. "Soln. dan." 1960 g. **6:** 66–71. (Interannual fluctuations of solar activity and negative anomalies of atmospheric circulation.)
13. Bezrukova, A. 1960. 11-letnii cykl solnetchnoi aktivnosti i kharakter kolebanii zemnoi zonalnoi circulatsii v zimnee vremia. Bull. "Soln. Dan." 1960 g. **7:** 78–82. (The 11-year cycle of solar activity and characteristics of the zonal circulation variations in winter time.)
14. Clayton, H. H. & F. L. Clayton. 1944 & 1947. World Weather Records. Smithsonian Misc. Collections, 79, 90, 105. Smithsonian Inst., Washington, D.C.

SOLAR, GEOMAGNETIC, AND METEOROLOGICAL PERIODICITIES

Fred Ward and Ralph Shapiro

Geophysics Research Directorate, United States Air Force, Cambridge Research Laboratories, Bedford, Mass.

Introduction

Tools now exist to determine precisely and objectively the existence of periodicities in time series (Blackman and Tukey, 1959). Since there still appears to be considerable controversy on the subject, as evidenced by the extensive literature and the comments in this monograph, it seems both desirable and illuminating to apply these tools in a comprehensive manner to a variety of astrogeophysical time series.

The astrogeophysical data used in this study are subjected to the procedures of generalized harmonic analysis. By these procedures it is possible to obtain the distribution of variance of a time series as a function of the frequency (or period) of the oscillations. Two sets of parameters have been chosen for study. One set is generally considered to represent the variable part of the solar energy output; the other represents variations of the large-scale meteorological circulations. The extent to which meteorological spectra reflect the characteristics of the solar spectra and the extent to which the atmosphere tends to show periodic variations will become evident from these results.

Data

Two essentially independent parameters have been chosen to represent solar variability. The first is the Zurich relative sunspot number R_z ; the second is the geomagnetic planetary 3-hour-range index K_p . The sunspot number can be considered to represent, to a moderate degree at least, variations of solar wave radiation and K_p , the variation of solar particle radiation.

The meteorological data consist of polynomial representations of the surface and 500-mb. (20,000-ft) pressure distributions, measures of the persistence of the surface-pressure distribution, indices describing the strength and character of the atmospheric circulation at the surface and at 500 mb., and the pressure and temperature at a fixed point on the earth's surface.

Analysis

FIGURE 1*a* presented for illustrative purposes, shows the spectrum of a random time series. The spectrum of a real time series shows a wide variety of oscillations, some of which may be real, while others are a result of sampling fluctuations. The spectrum of a series of random numbers shows variations resulting only from sampling fluctuations. The spectrum of a series of random numbers is commonly referred to as "white noise"; that is, it has equal variance (power) at all frequencies.

The ordinate in FIGURE 1*a* is the fraction of the total variance of the time series (U_h). The abscissa is the frequency (h) on a linear scale; the period scale corresponding to these frequencies is indicated above the data points.

The period and frequency are related to each other as a function of the number of lags for which the autocorrelation function is computed. The period times the frequency equals 2M, where M is the number of lags. The highest frequency is equal to M, and corresponds to a period that is twice the sampling interval. The total area under the curve, or the sum of all of the data points in the spectrum, has been made equal to unity. This normalization of the data makes the various spectra more easily intercomparable. The most obvious feature of FIGURE 1*a* is that the level of the continuum is constant. There are, however, many peaks in the spectrum of which approximately 5 per cent are significant at the 5-per cent level. It should be noted that the level of the continuum in FIGURE 1*a* is 0.0025.

FIGURE 1*b* is the spectrum of the daily Zurich relative sunspot number for the period April 1, 1951, through March 31, 1956, a total of 1826 data points. This period corresponds to a time around sunspot minimum. There are 2 noteworthy features of this figure. The first is the broad maximum corresponding to the 27-day solar-rotation period. The other is the extremely high values of the variance near h = 0 (noted by the arrows with the corresponding numerical values of the variance). These correspond to periods that are too long to be resolved with a sample of this size; however, they are probably due to the 11-yr. sunspot cycle.

FIGURE 2*a* shows the spectrum of the sunspot number for 2 different time periods. The bottom part of FIGURE 2*a* is for the same time period as that of FIGURE 1*b*, whereas the top part is for a different time period in roughly the same phase of the sunspot cycle. M in FIGURE 2*a* is equal to 100, contrasted with an M of 400 in FIGURES 1*a* and *b*. Hence the resolution of this figure is one quarter of that of the previous 2 figures. Both of these spectra are very similar, each showing a pronounced peak around 27 days, and also considerable variance corresponding to a long-period oscillation (around h = 0).

FIGURE 2*b* shows the spectrum of the central zone Zurich relative sunspot number. Since large spot groups are usually visible on the disc for 14 consecutive days, the whole disc-sunspot number (R_z) can be considered to be a 14-day smoothed parameter. Hence variations of 14 days and less are almost completely damped out. The central zone sunspot number takes into account only sunspots that are in a central circle whose diameter is the solar radius. A time series of central-zone numbers is roughly equivalent to a 5-day smoothed quantity. The time period covered by FIGURE 2*b* is the same as that in the top part of FIGURE 2*a*. It shows not only the 27-day period and the longer period variations near h = 0, but also a pronounced maximum around 14 days. This maximum could be interpreted as a harmonic of the fundamental rotation period, occurring at 27/2 days. Alternatively it might be interpreted as evidence of a frequent tendency for 2 spot groups to exist simultaneously, separated by approximately 180° of solar longitude.

FIGURE 3 is the spectrum of a time series of monthly averages of the Zurich relative sunspot number. The most striking feature of this spectrum is the very large fraction of the variance explained by the 11-yr. sunspot period, amounting to more than one half the total variance of the time series. The small maximum at h = 12 corresponds to a period of 5½ yr., and most likely is due to the fact that the shape of the 11-yr. variation does not correspond ex-

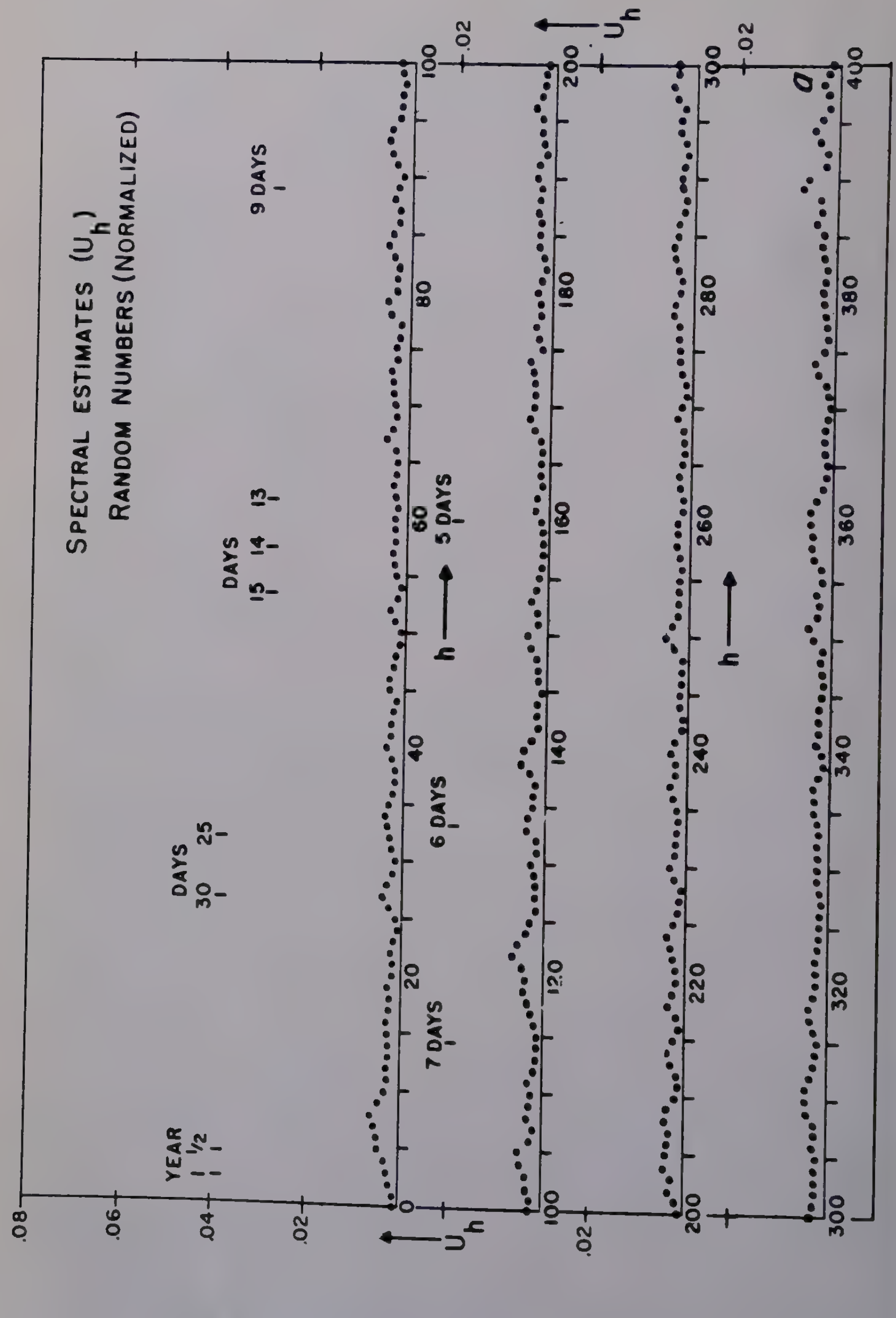

SPECTRAL ESTIMATES (U_h)
RANDOM NUMBERS (NORMALIZED)
YEAR
1 1/2
DAYS
30 25
DAYS
15 14 13
9 DAYS
7 DAYS
6 DAYS
5 DAYS
h
U_h
a

FIGURE 1(a).

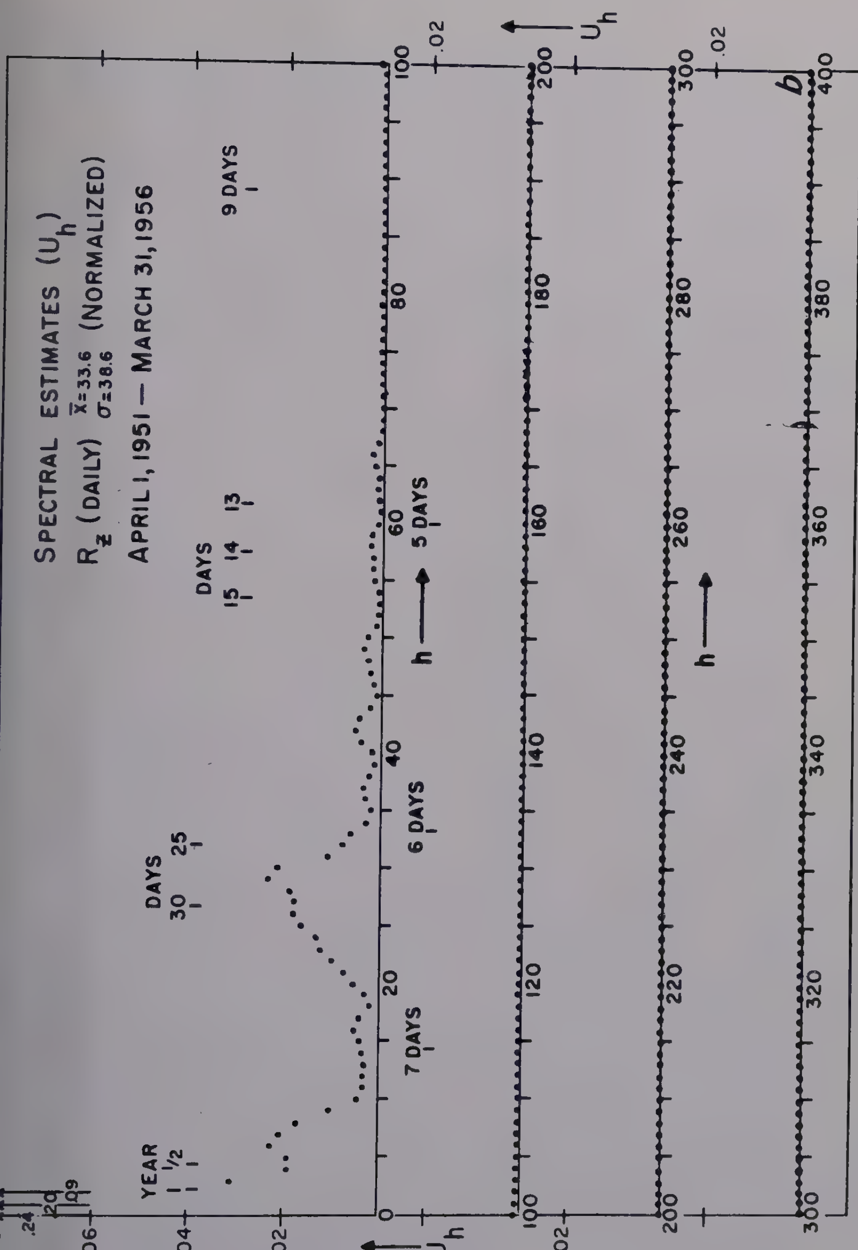
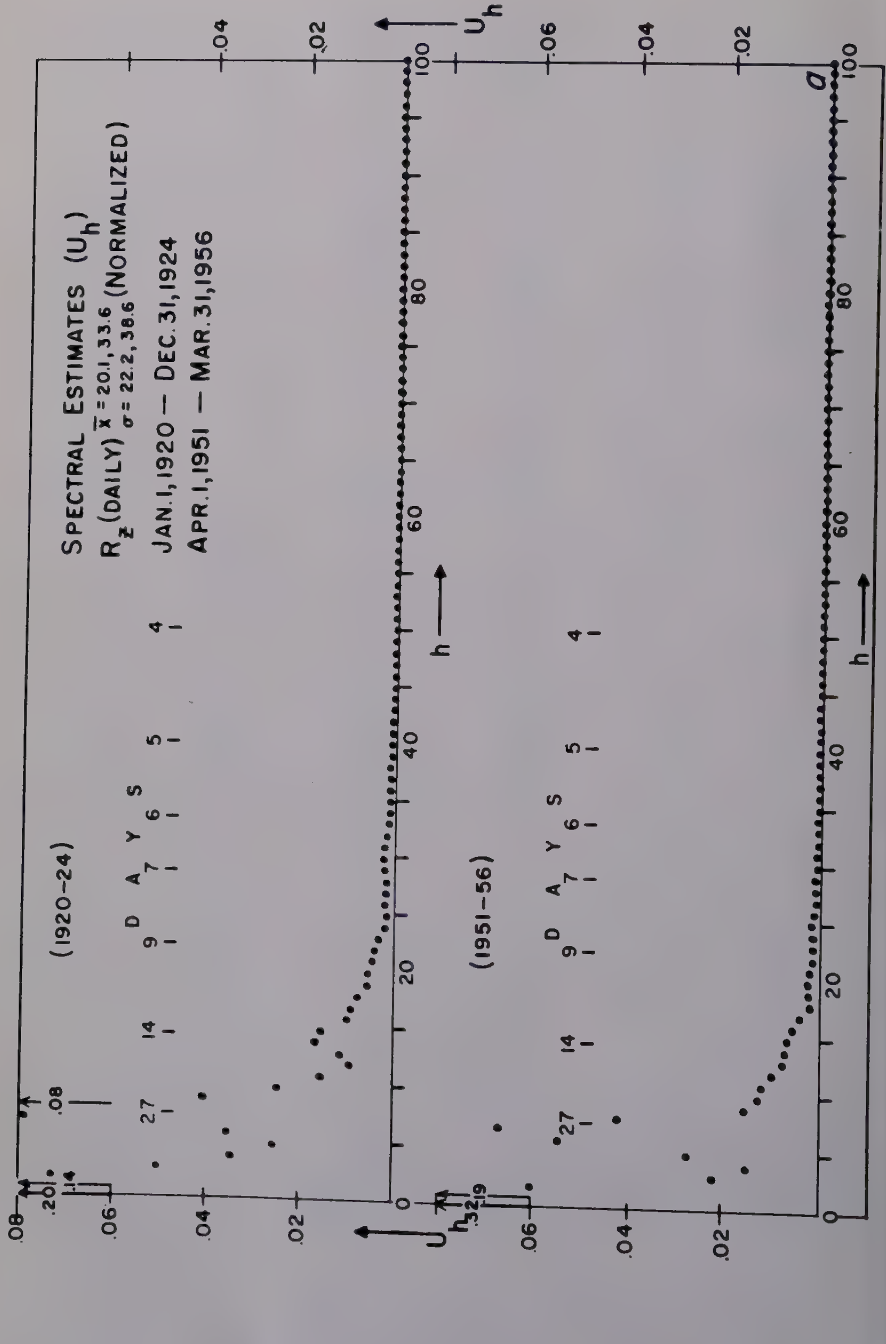


FIGURE 1. (*a*) Variance spectrum of random numbers. (*b*) Variance spectrum of daily Zurich relative sunspot number (R_z).



SPECTRAL ESTIMATES (U_h)
R_z (DAILY) x̄ = 20.1, 33.6 σ = 22.2, 38.6 (NORMALIZED)
JAN.1,1920 — DEC.31,1924
APR.1,1951 — MAR.31,1956
(1920-24)
(1951-56)
DAYS
27 14 9 7 6 5 4
h
U_h
a

FIGURE 2(a).

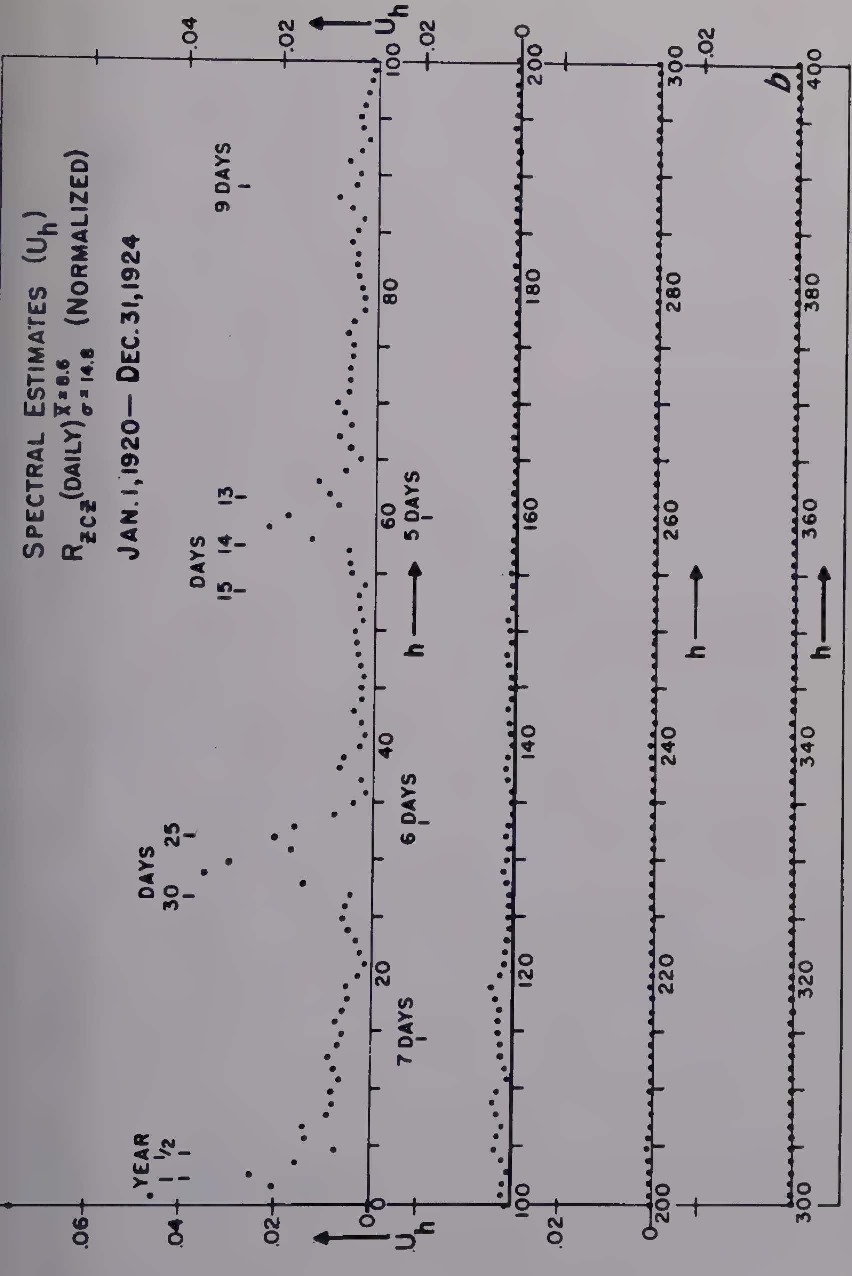


FIGURE 2. (*a*) Variance spectra of daily Zurich relative sunspot number (R_z). (*b*) Variance spectrum of daily central-zone Zurich relative sunspot number (R_{zcz}).

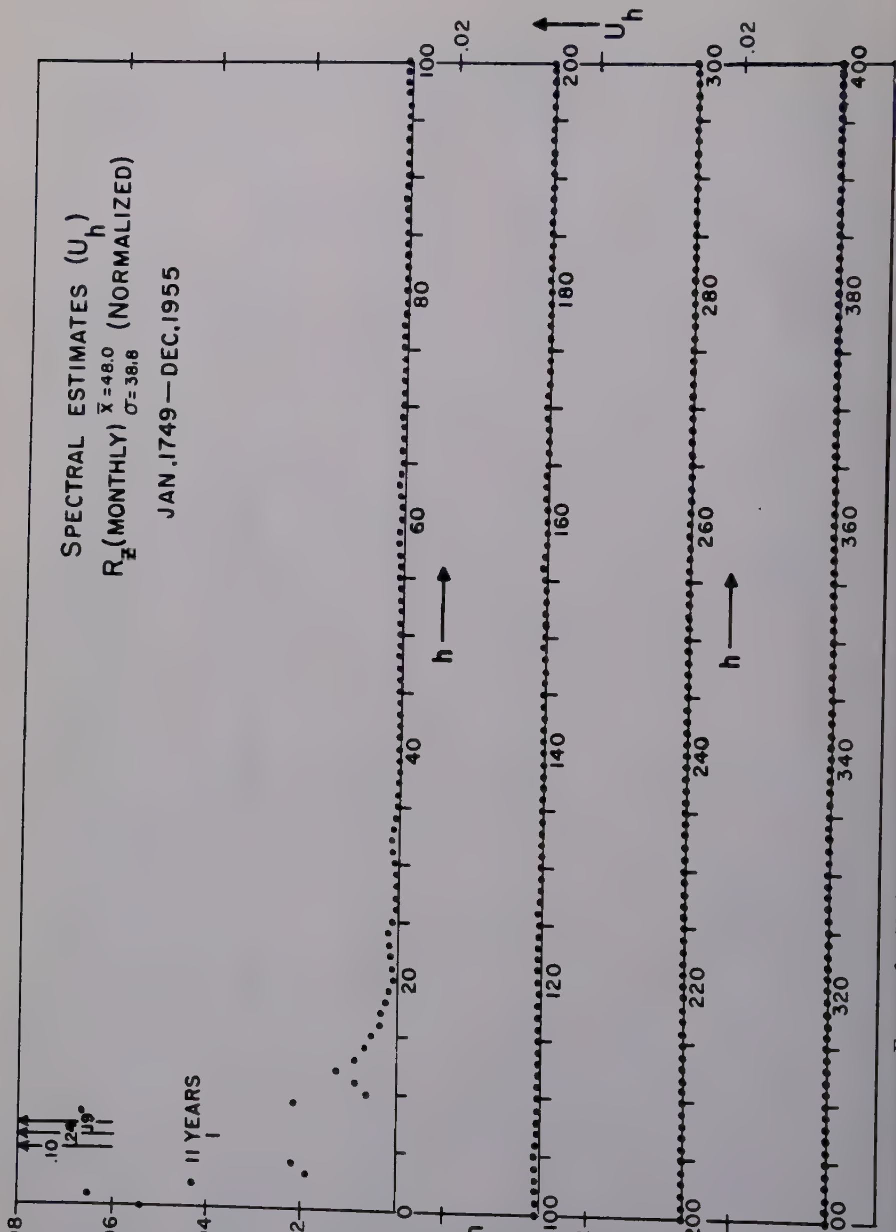


FIGURE 3. Variance spectrum of monthly mean Zurich relative sunspot number (R

actly to a sine wave. Resolution of significant variations in the spectrum at periods longer than 11 yr. is very difficult. As a matter of interest however at h = 3, corresponding to the 22-yr. period, there is a minimum in the spectrum.

FIGURE 4 is the spectrum of a time series of 5 yr. of daily values of the geomagnetic planetary 3-hour-range index K_p. This period of time is the same as that of FIGURE 1*b* and occurs around sunspot minimum. This spectrum shows maxima corresponding to 6-month and 27-day periods. These maxima account for roughly 6 and 16 per cent of the total variance of the time series respectively. These periods are well known and require no further comment. However there are also maxima corresponding to the first 4 harmonics of the 27-day period (14, 9, 7, and 5½ days). Analogous to FIGURE 2*b*, these harmonics can be interpreted either as distortions of the basic 27-day solar-rotation period or as tendencies toward equidistant spacing of solar-corpuscular emitting regions.

FIGURE 5 is the spectrum of K_p for a different 5-yr. period, corresponding to a time around sunspot maximum. This spectrum clearly shows the 6-month and 27-, 14-, and 9-day periods, but with amplitudes different from those in FIGURE 4. The 7- and 5½-day periods are not distinguishable from the background noise in FIGURE 5.

The remaining figures show spectra for a variety of large-scaled meteorological parameters. FIGURE 6 shows the spectra*(for 4 separate 5-yr. periods) of the surface zonal westerlies for a European region (Z_{E7}) and a North American region (Z_{N7}). These zonal westerlies were determined from the surface-pressure distribution by orthogonal-polynomial representation. The European region is bounded by lat. 35 and 65 N, and lon. 30 E and 30 W. The North American region is bounded by lat. 30 and 60 N, and lon. 65 W and 125 W. The top 4 spectra in FIGURE 6 are for time periods around sunspot maximum; the lower 4 are around sunspot minimum. Although there exist standard tests for the determination of the statistical significance of a spectral peak, consistency of spectral variations during independent time periods generally is considered to be a more reliable measure of significance. All 8 spectra in this figure show an annual period. In addition, they show scattered maxima. There are however no consistent maxima for the same index among the different spectra, except for the annual period. An interesting feature in all of these spectra is the decrease in the level of the continuum with increasing frequency (decreasing period). This "red noise" in the spectra is analogous to persistence in the time domain, and reflects the fact that most large-scale meteorological parameters have appreciable autocorrelation. Although it is not as readily apparent, FIGURES 1*b* through 5 also exhibit this same phenomenon.

FIGURE 7 contains the spectra of the same quantities as in FIGURE 6 for the same time periods, but with M equal to 100. These lower resolution spectra (one fourth of the resolution of FIGURE 6) illustrate the fluctuations in the range from 2 days to approximately 1 mo. more readily than those in FIGURE 6. It is apparent that the only noteworthy and consistent feature in the spectra shown in FIGURE 7 is the red-noise phenomenon.

* Only the first 100 points of these high-resolution spectra are presented. Therefore the shortest period shown is 8 days.

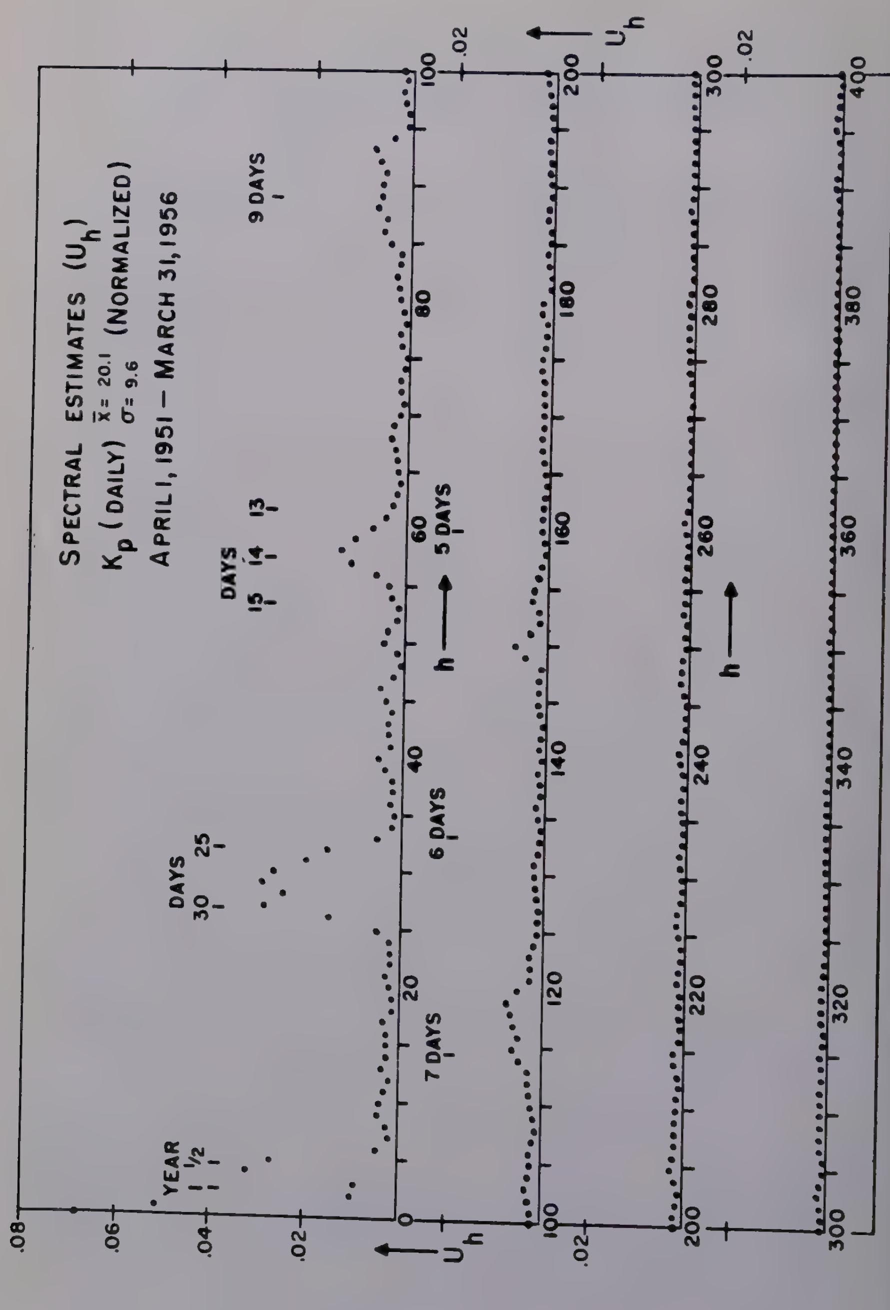


FIGURE 4. Variance spectrum of geomagnetic planetary 3-hour-range index (K_p).

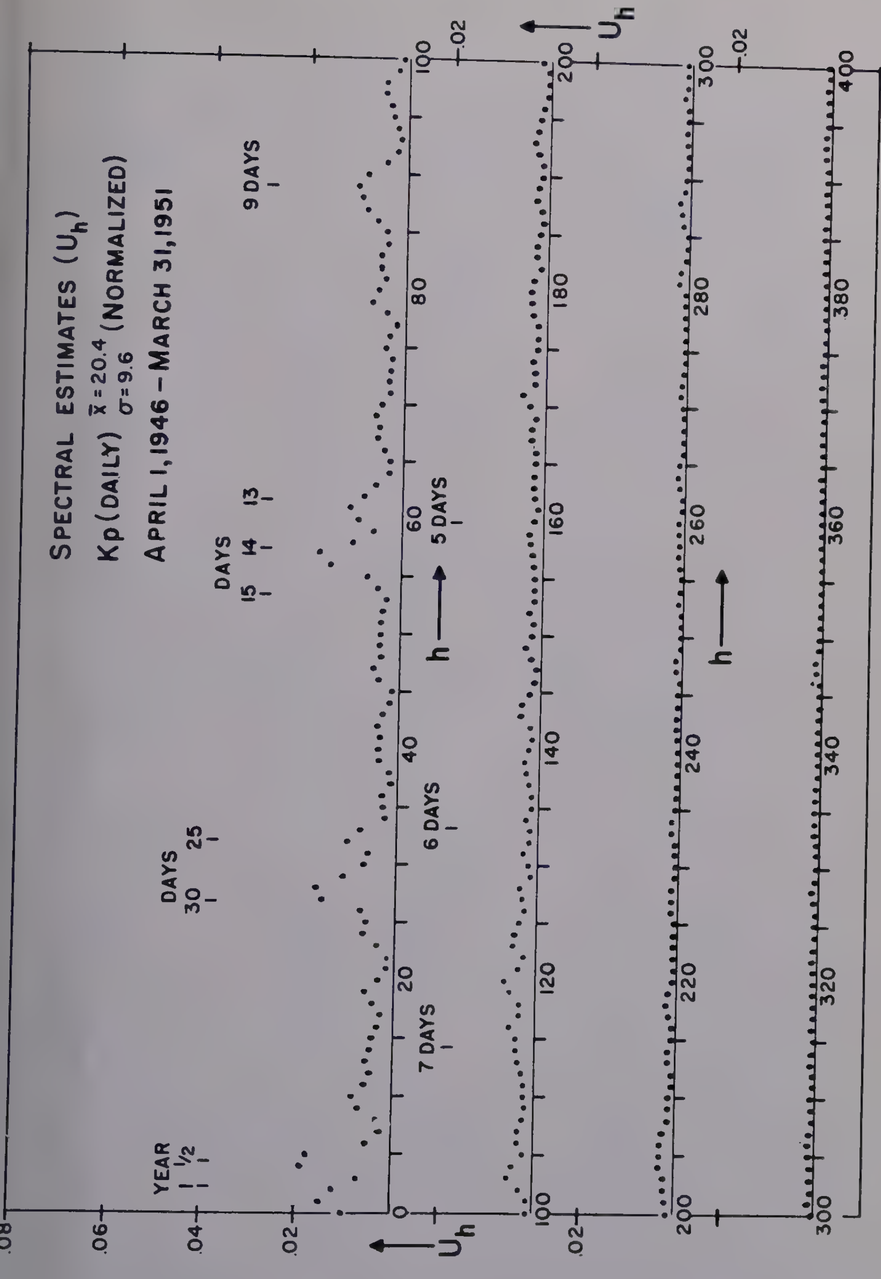
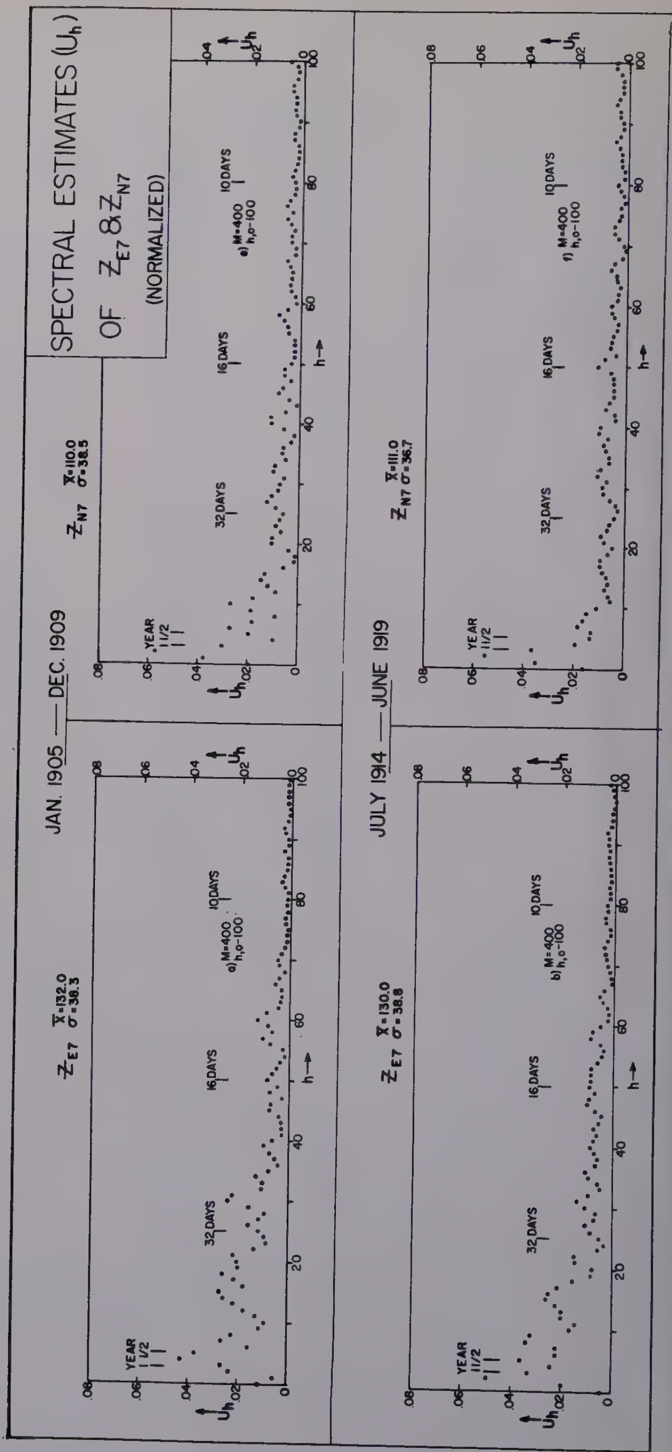


FIGURE 5. Variance spectrum of geomagnetic planetary 3-hour-range index (K_p).



SPECTRAL ESTIMATES (U_h)
OF Z_{E7} & Z_{N7}
(NORMALIZED)
JAN. 1905 — DEC. 1909
JULY 1914 — JUNE 1919
Z_{E7} $\bar{X}=132.0$ $\sigma=38.3$
Z_{N7} $\bar{X}=110.0$ $\sigma=38.5$
Z_{E7} $\bar{X}=130.0$ $\sigma=38.8$
Z_{N7} $\bar{X}=111.0$ $\sigma=36.7$
a) $M=400$ $h,0-100$
e) $M=400$ $h,0-100$
b) $M=400$ $h,0-100$
f) $M=400$ $h,0-100$
YEAR
1 1/2
32 DAYS
16 DAYS
10 DAYS
U_h
$h\rightarrow$

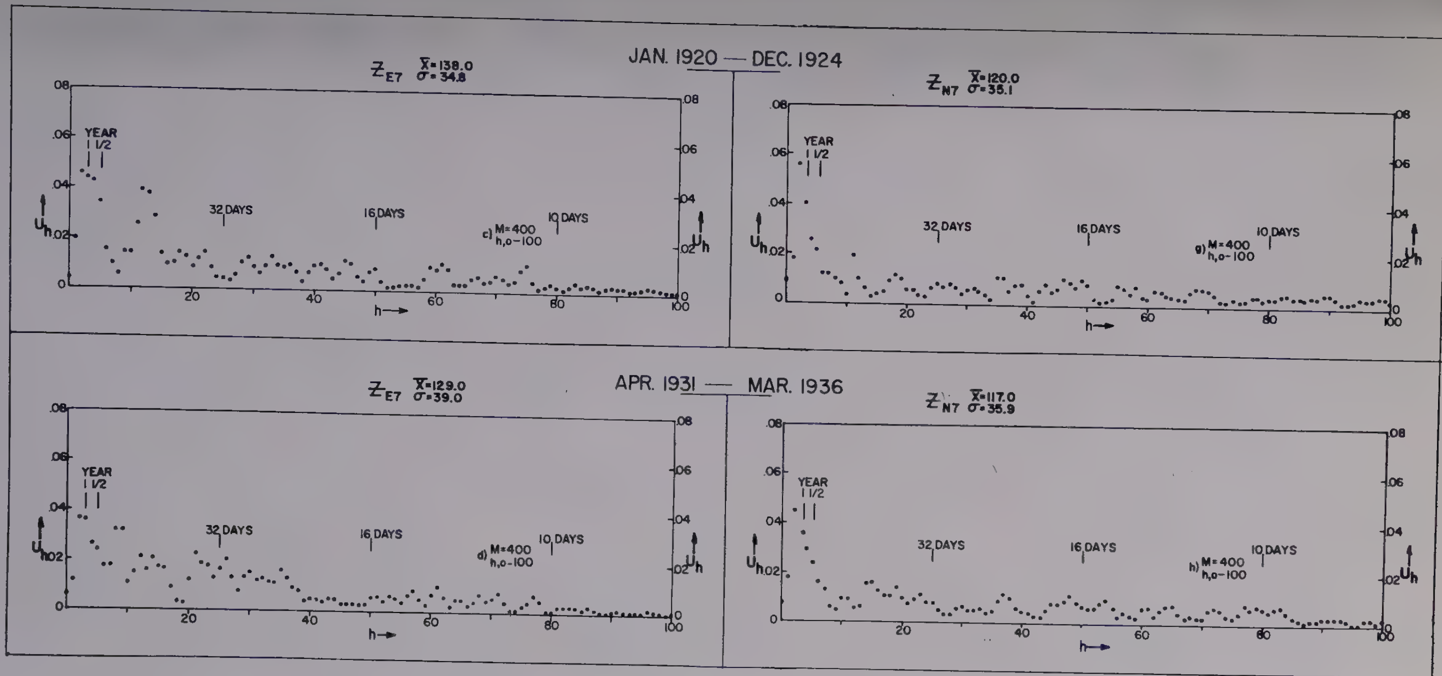
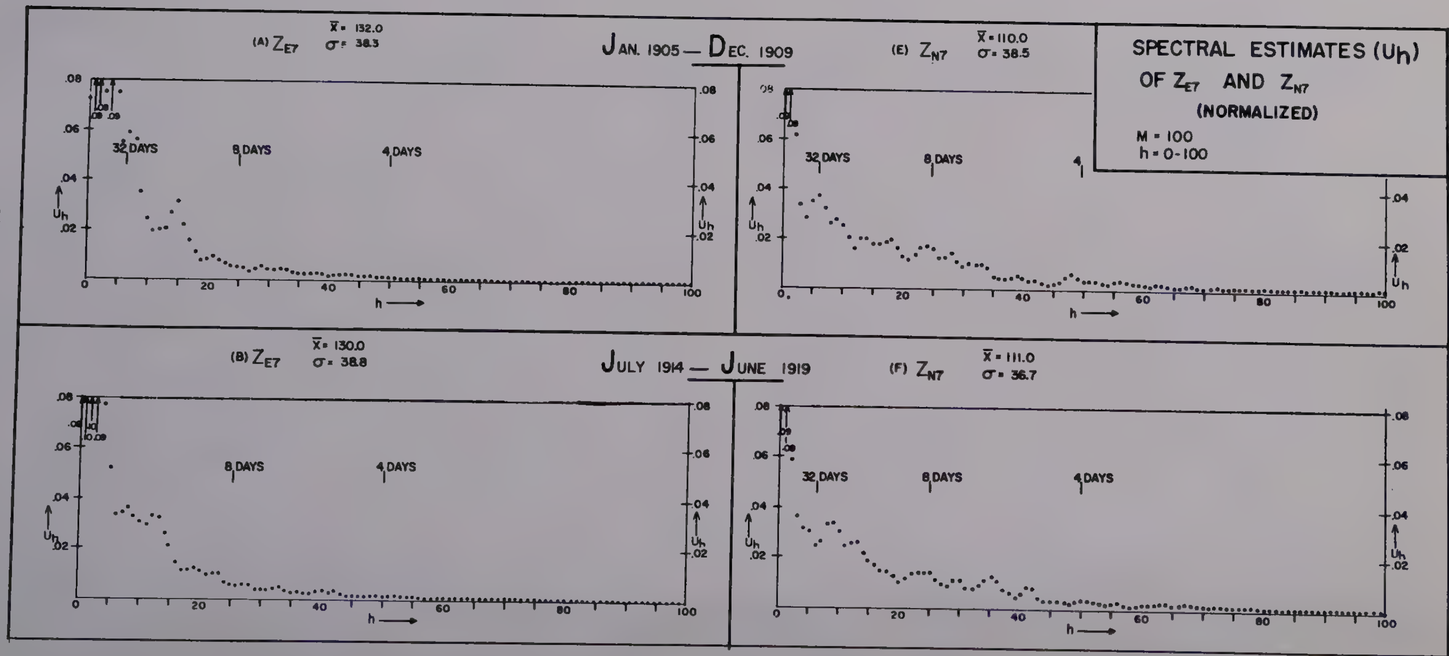


FIGURE 6. Variance spectra of surface zonal winds (Z_{E7}, Z_{N7}).


SPECTRAL ESTIMATES (Uh) OF ZE7 AND ZN7
(NORMALIZED)
M = 100
h = 0-100
JAN. 1905 — DEC. 1909
(A) ZE7
X̄ = 132.0
σ = 38.3
(E) ZN7
X̄ = 110.0
σ = 38.5
JULY 1914 — JUNE 1919
(B) ZE7
X̄ = 130.0
σ = 38.8
(F) ZN7
X̄ = 111.0
σ = 36.7
32 DAYS
8 DAYS
4 DAYS
Uh
h

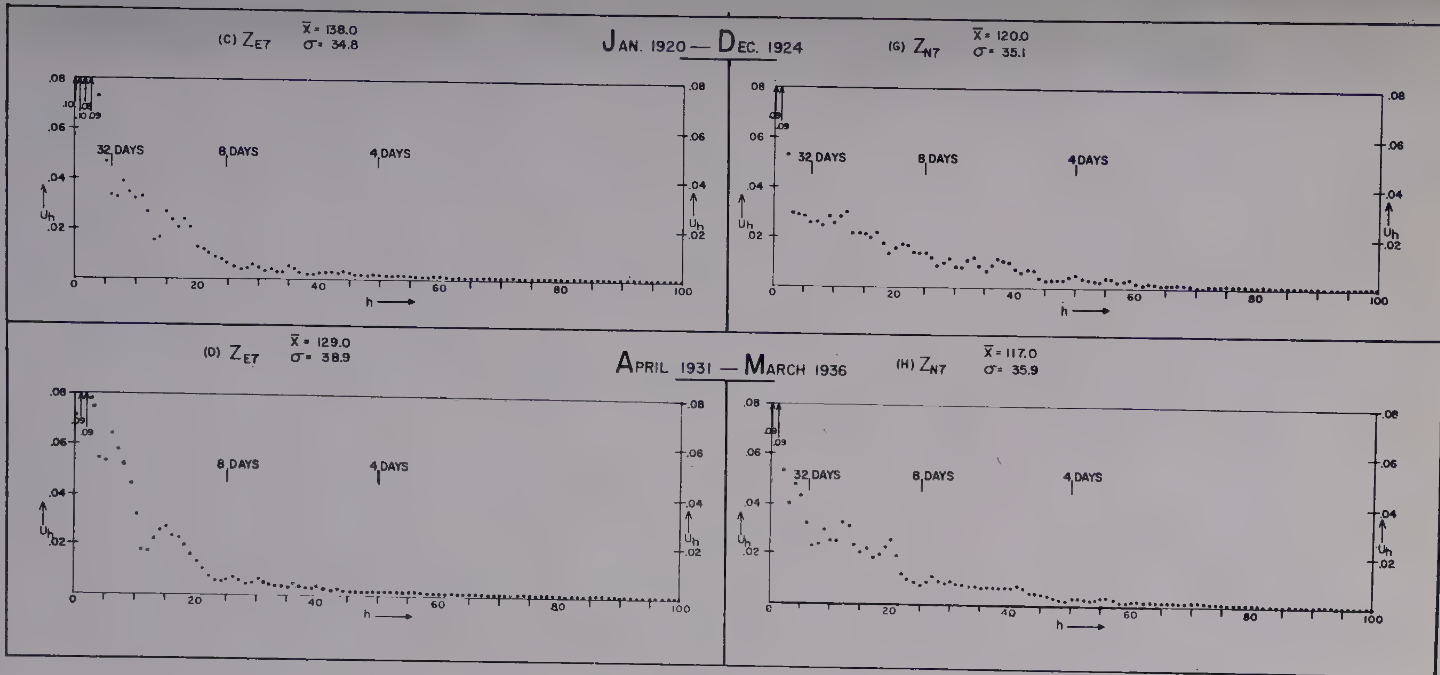
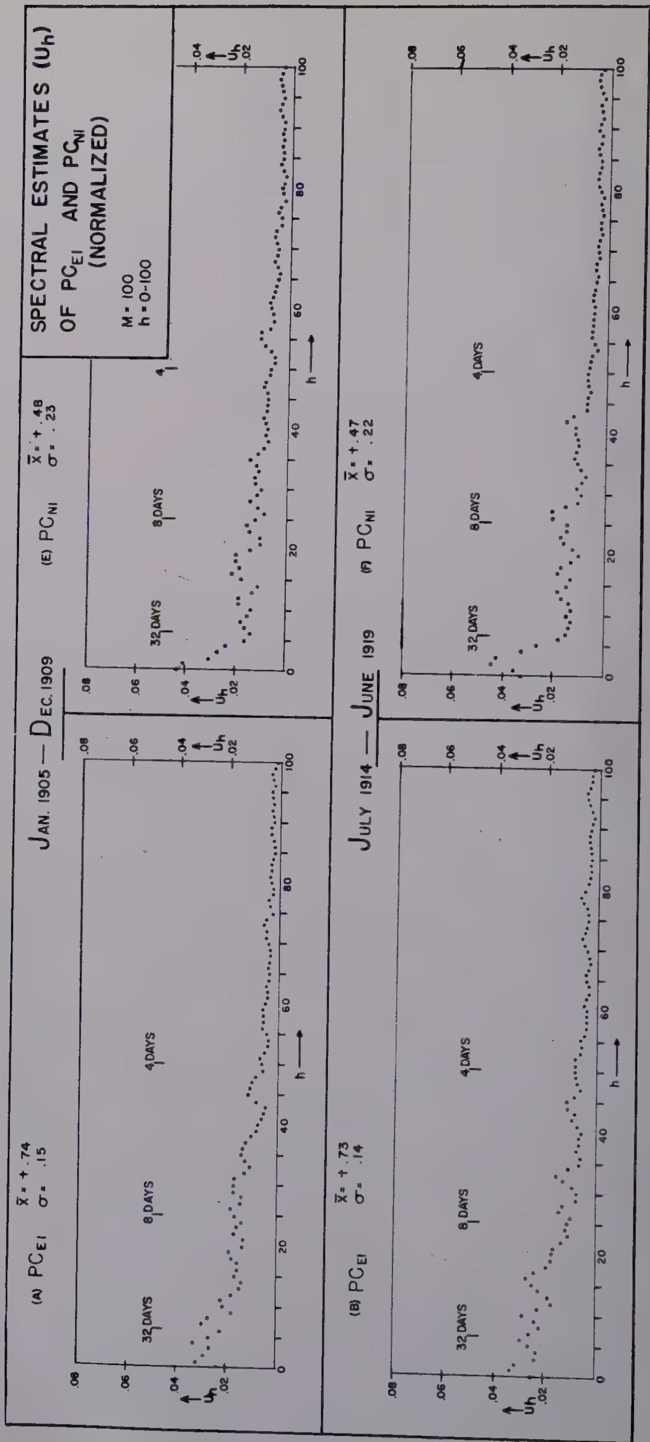


FIGURE 7. Variance spectra of surface zonal winds (Z_{E7}, Z_{N7}).


SPECTRAL ESTIMATES (U_h) OF PC_{EI} AND PC_{NI} (NORMALIZED)
M = 100
h = 0-100
JAN. 1905 — DEC. 1909
(A) PC_{EI}
$\bar{X}$ = +.74
σ = .15
(E) PC_{NI}
$\bar{X}$ = +.48
σ = .23
JULY 1914 — JUNE 1919
(B) PC_{EI}
$\bar{X}$ = +.73
σ = .14
(F) PC_{NI}
$\bar{X}$ = +.47
σ = .22
32 DAYS
8 DAYS
4 DAYS
U_h
h
.08
.06
.04
.02
0
20
40
60
80
100

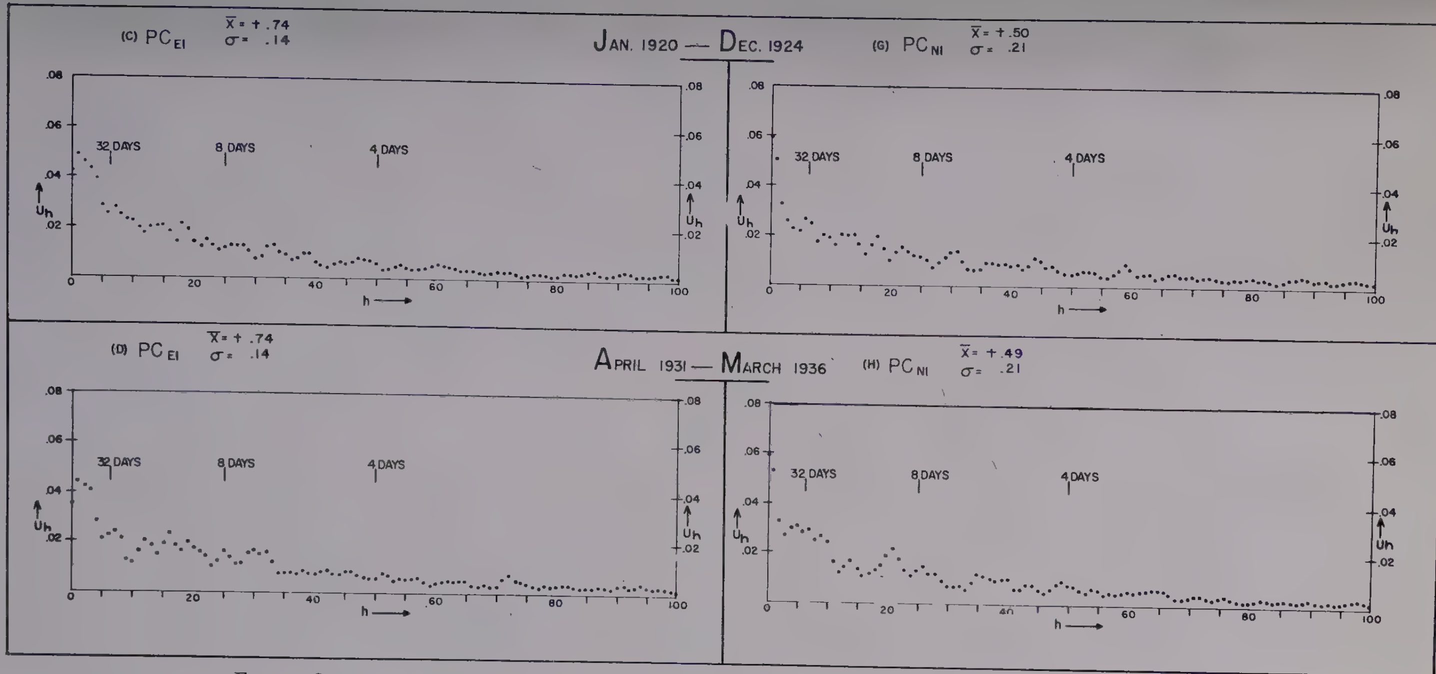
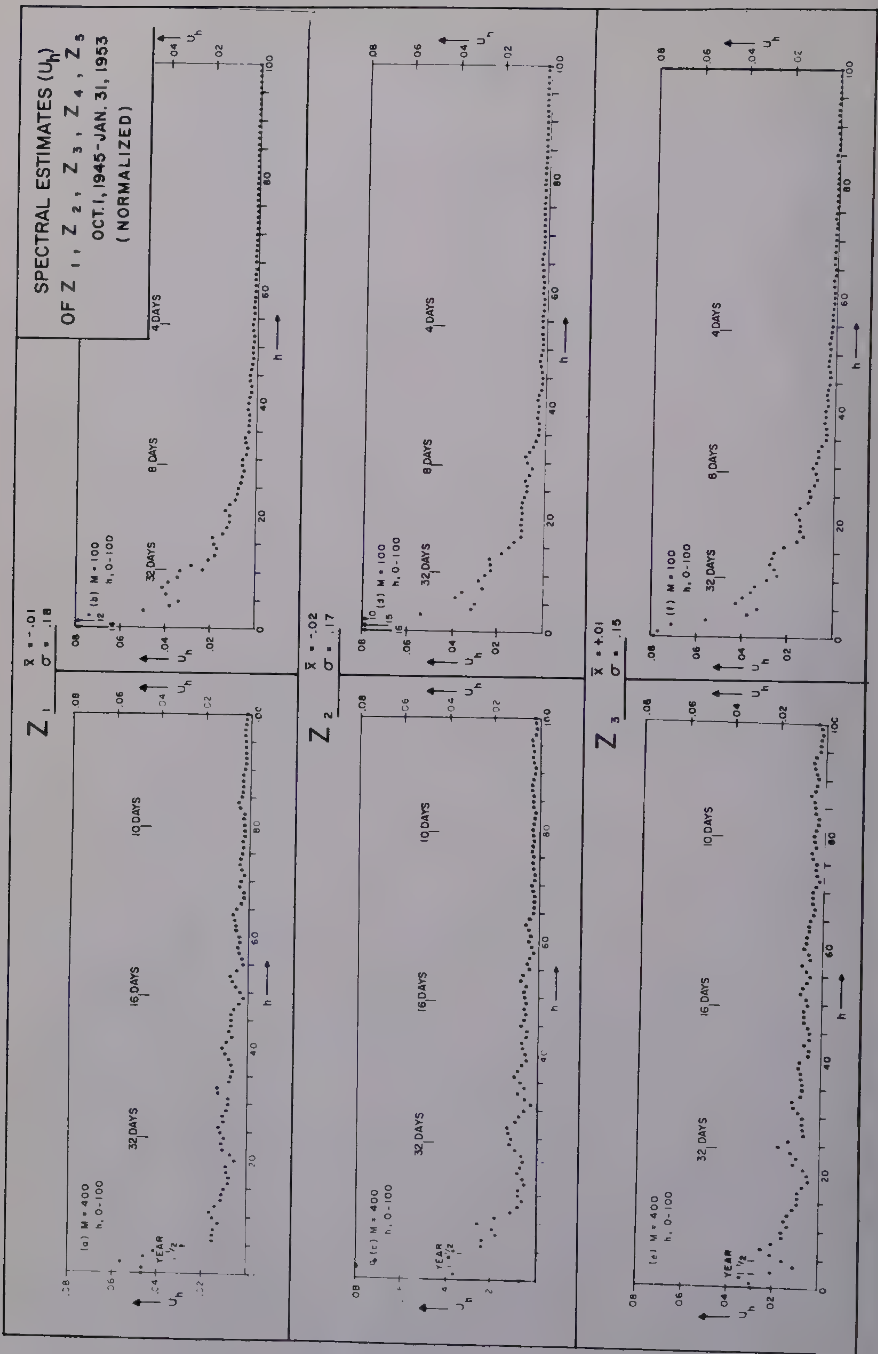


FIGURE 8. Variance spectra of one-day lag, sea-level pressure persistence correlations (PC_{EI}, PC_{NI}).


SPECTRAL ESTIMATES (U_h)
OF Z_1, Z_2, Z_3, Z_4, Z_5
OCT. 1, 1945-JAN. 31, 1953
(NORMALIZED)
Z_1
$\bar{X}$ = -.01
σ = .18
Z_2
$\bar{X}$ = -.02
σ = .17
Z_3
$\bar{X}$ = +.01
σ = .15
(a) M = 400
h, 0-100
(b) M = 100
h, 0-100
(c) M = 400
h, 0-100
(d) M = 100
h, 0-100
(e) M = 400
h, 0-100
(f) M = 100
h, 0-100
YEAR
32 DAYS
16 DAYS
10 DAYS
8 DAYS
4 DAYS
U_h
h

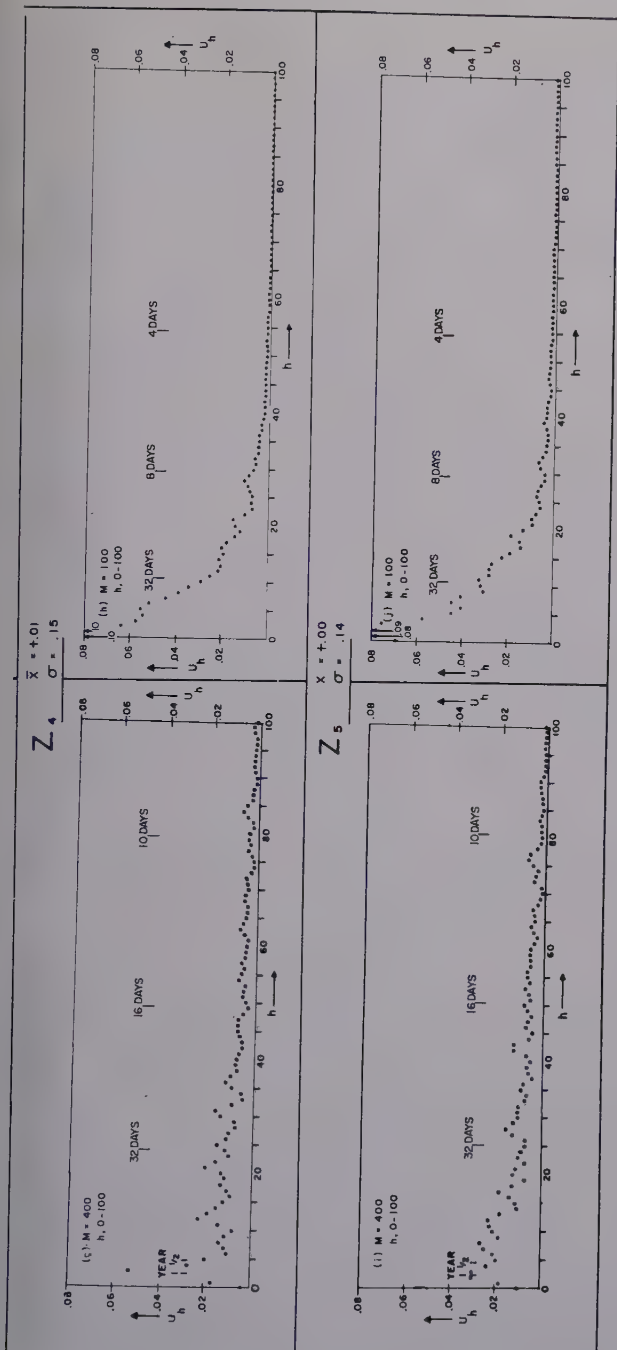
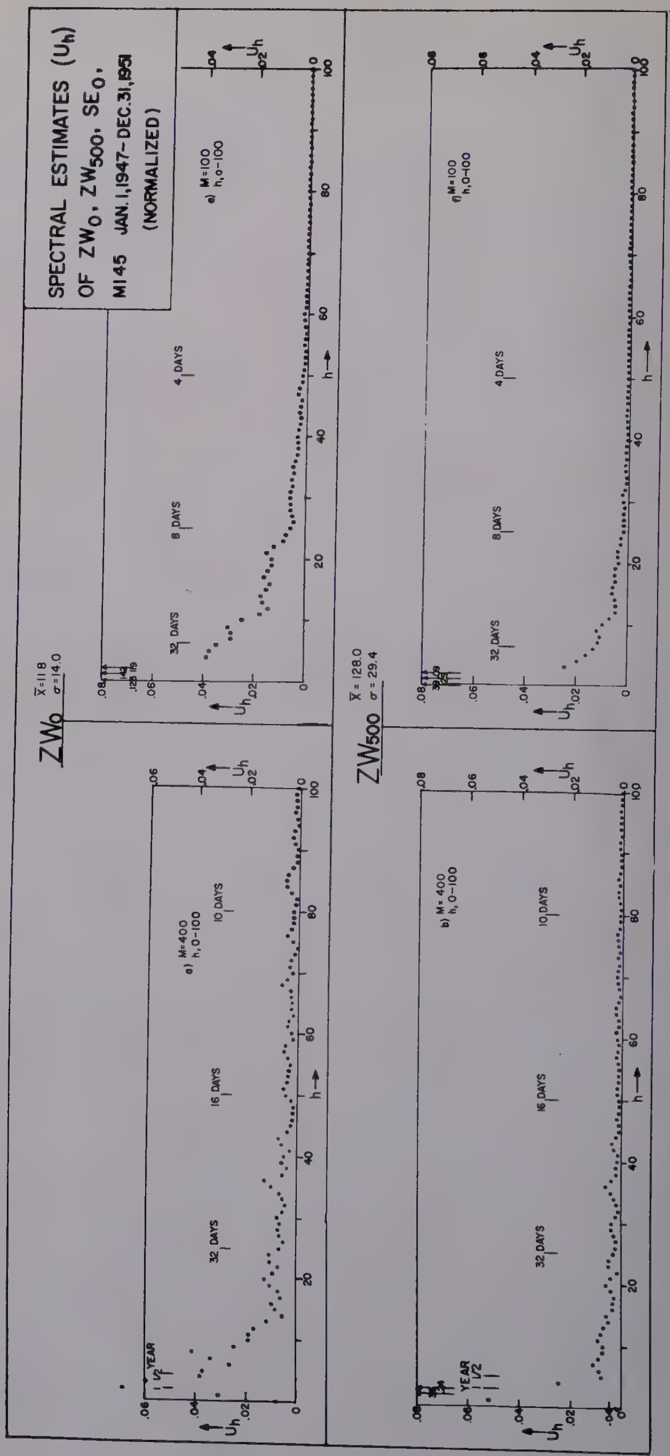


FIGURE 9. Variance spectra of 500-millibar polynomials (Z_1, Z_2, Z_3, Z_4, Z_5).



SPECTRAL ESTIMATES (Uh)
OF ZW0, ZW500, SE0,
M145 JAN. 1, 1947–DEC. 31, 1951
(NORMALIZED)
ZW0
X̄ = 11.8
σ = 14.0
ZW500
X̄ = 128.0
σ = 29.4
a) M=400 h, 0–100
b) M=400 h, 0–100
e) M=100 h, 0–100
f) M=100 h, 0–100
YEAR
1/2
32 DAYS
16 DAYS
10 DAYS
8 DAYS
4 DAYS
Uh
h

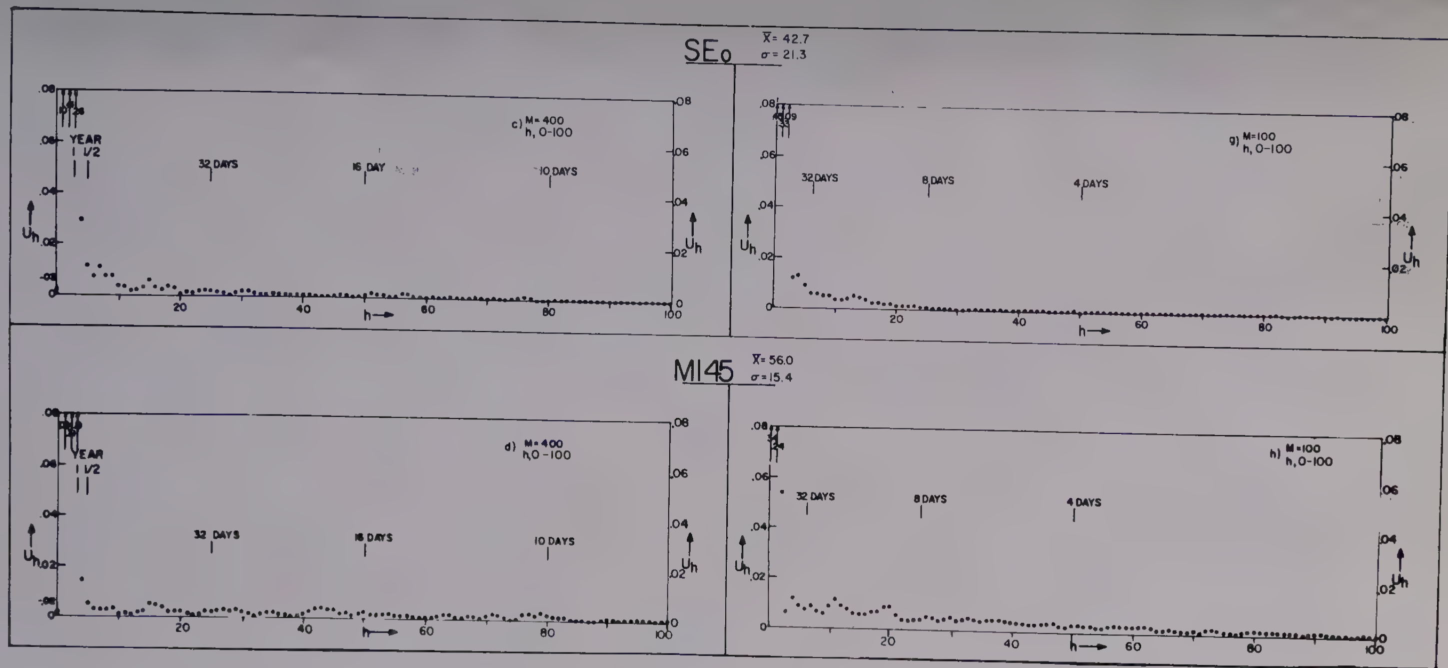
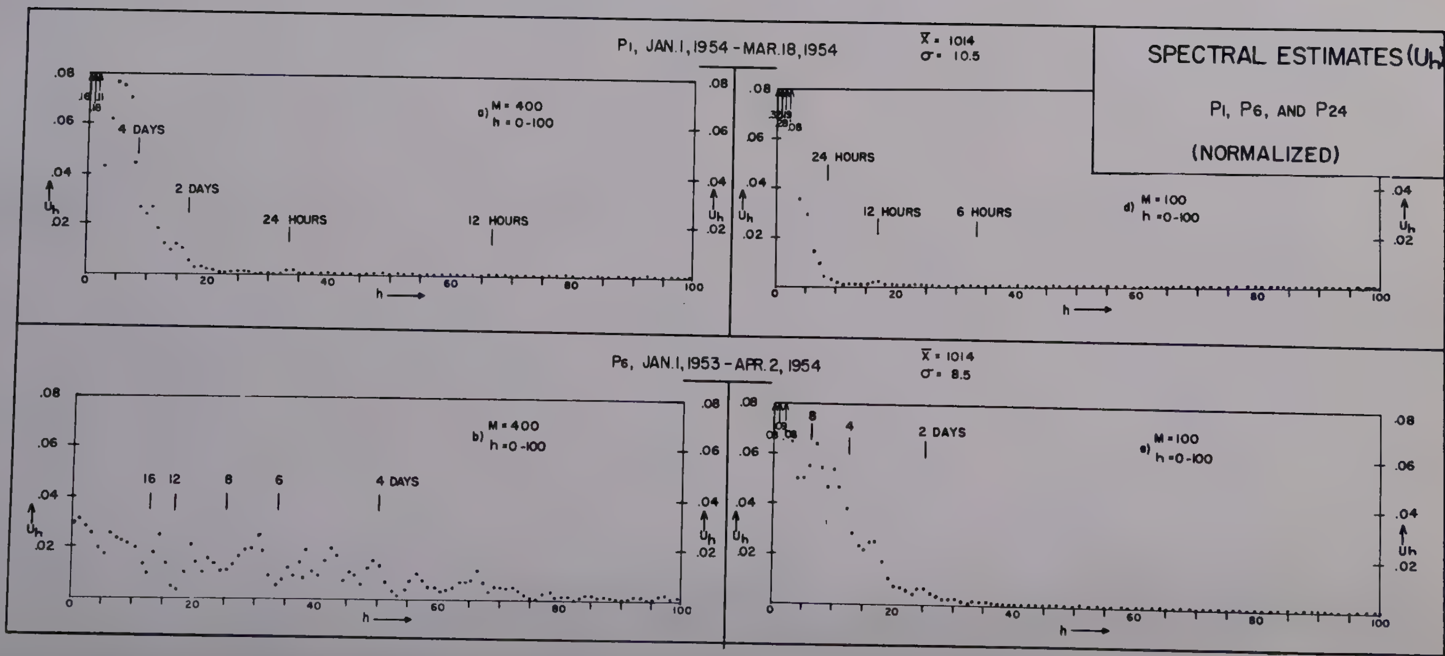


FIGURE 10. Variance spectra of surface and 500-mb. hemispheric circulation indices (ZW_0, ZW_{500}, SE_0, MI45).



SPECTRAL ESTIMATES (U_h)
P1, P6, AND P24
(NORMALIZED)
P1, JAN. 1, 1954 - MAR. 18, 1954
$\bar{X}$ = 1014
σ = 10.5
a) M = 400
h = 0-100
4 DAYS
2 DAYS
24 HOURS
12 HOURS
d) M = 100
h = 0-100
24 HOURS
12 HOURS
6 HOURS
P6, JAN. 1, 1953 - APR. 2, 1954
$\bar{X}$ = 1014
σ = 8.5
b) M = 400
h = 0-100
16
12
8
6
4 DAYS
e) M = 100
h = 0-100
8
4
2 DAYS
U_h
h

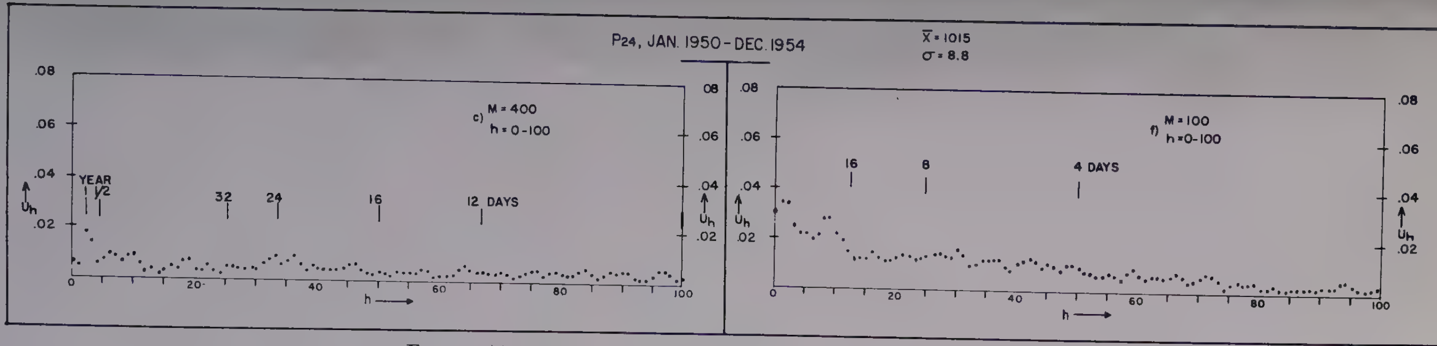
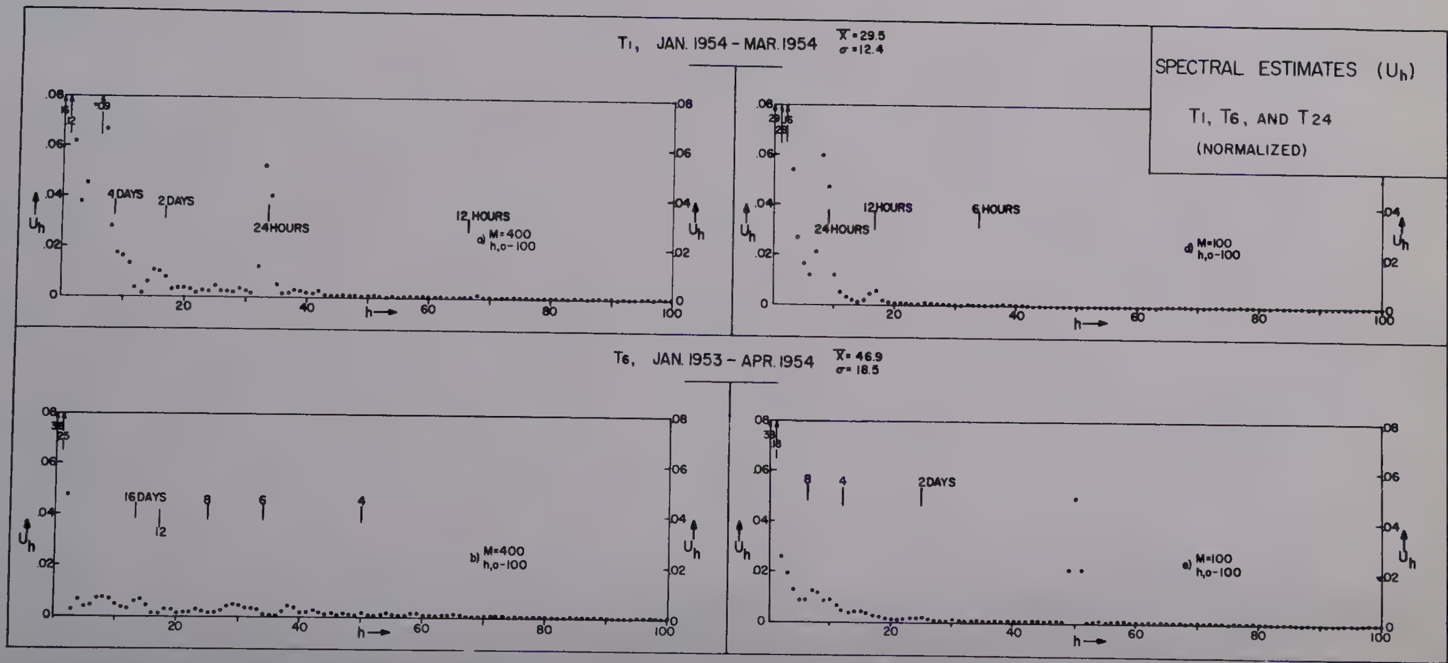


FIGURE 11. Variance spectra of surface pressure at Bedford, Mass.


SPECTRAL ESTIMATES (U_h)
T_1, T_6, AND T_{24}
(NORMALIZED)
T_1, JAN. 1954 – MAR. 1954
$\overline{X} = 29.5$
$\sigma = 12.4$
4 DAYS
2 DAYS
24 HOURS
12 HOURS
a) M=400 h,o-100
12 HOURS
24 HOURS
6 HOURS
d) M=100 h,o-100
T_6, JAN. 1953 – APR. 1954
$\overline{X} = 46.9$
$\sigma = 18.5$
16 DAYS
12
8
6
4
b) M=400 h,o-100
8
4
2 DAYS
e) M=100 h,o-100
U_h
$h \rightarrow$

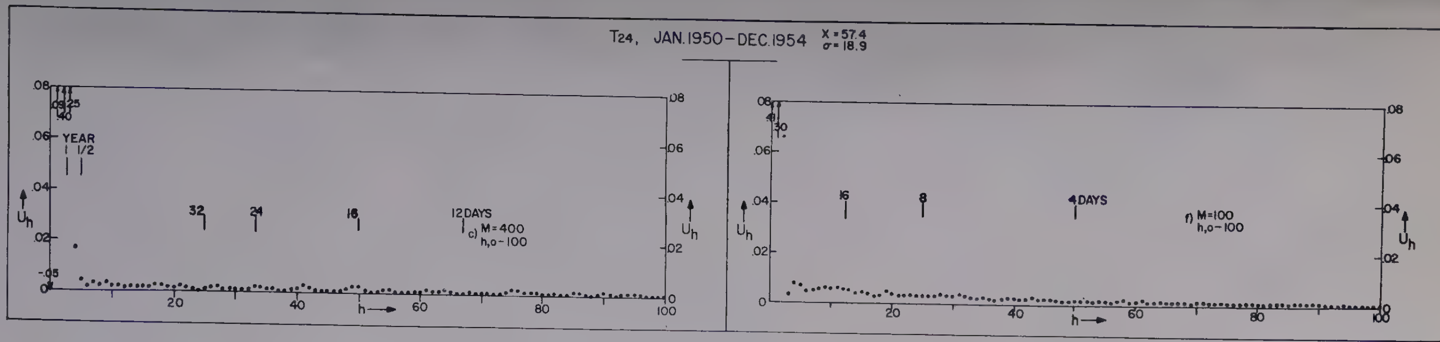


FIGURE 12. Variance spectra of surface temperature at Bedford, Mass.

FIGURE 8 shows spectra (for the same 4 time periods as in FIGURES 6 and 7) of the 1-day lag persistence correlation of the daily surface-pressure distribution over each of the 2 areas. This figure shows only the low-resolution spectra ($M = 100$) and contains excellent examples of "red noise."

FIGURE 9 contains the spectra of the first 5 of a set of 22 orthogonal polynomials computed to fit a one-half-hemisphere, 500-mb. topography. Z_1 through Z_5 are respectively the first- through fifth-degree representations in the east-west direction, and represent the long waves. The 5 diagrams on the left in this figure contain the first 100 points of the high-resolution spectra ($M = 400$); the 5 figures on the right contain the entire low-resolution spectra ($M = 100$). Z_1, Z_2, and Z_4 show an obvious annual period. All however have a red-noise continuum with superimposed sampling fluctuations.

FIGURE 10 depicts the spectra of 4 hemispherical circulation indices. The diagrams on the left have higher resolution ($M = 400$) than those on the right ($M = 100$). The annual period is very distinct in all high-resolution spectra. The annual period is least pronounced in the surface zonal westerlies (ZW_0) where it accounts for about 10 per cent of the total variance, and most pronounced in the subtropical easterlies (SE_0) where it explains about two thirds of the total variance. The only other characteristic of these spectra is their resemblance to red noise.

FIGURES 11 and 12 are the spectra of the pressure and temperature recorded at the weather station at Hanscom Field, Bedford, Mass. It was possible with these data to extend the analysis of the short-period end of the spectrum to 2 hours. The detail of the spectrum of the hourly values shows the diurnal and semidiurnal waves which, however, are minute compared to the variations over periods of many days. The hourly temperature spectra show the diurnal period very distinctly and, in addition, a small peak at 12 hours that undoubtedly is due to distortion of the 24-hour wave. The high-resolution spectra of the hourly pressures and temperatures also show maxima at $h = 5$ ($6\frac{2}{3}$ days) and $h = 6$ ($5\frac{1}{2}$ days) respectively. However, such periods are not shown in the 6-hourly or daily spectra that are computed for much longer records (same number of data points), and hence provide more reliable estimates. It must be concluded therefore that the occurrence of these periods is due to statistical sampling fluctuations.

Conclusions

On the basis of the results of this study, it is possible to arrive at the following general conclusions. Sunspot numbers are persistent and show the 11-yr. and 27-day periods, and harmonics of the 27-day period. K_p (and presumably solar-particle radiation) is persistent and shows periods at 11 yr. (not shown), 6 mo., 27 days, and harmonics of the 27-day period. Meteorological data are persistent and show annual, diurnal, and semidiurnal periods.

Since the persistence characteristic is the only similarity in both the solar and meteorological data, it would seem that for many purposes the astronomers and meteorologists are sampling data too frequently.

Reference

BLACKMAN, R. B. & J. W. TUKEY. 1959. The Measurement of Power Spectra, 190 pp. Dover. New York, N.Y.

PHYSICAL ASPECTS OF DEDUCED AND ACTUAL CLIMATIC CHANGE

E. B. Kraus
*Woods Hole Oceanographic Institution, Woods Hole, Mass.**

Introduction

The general circulation is driven by differential heating, but the distribution of the heat sources and sinks depends on the circulation in turn. The system is complex, and allows a large number of different transient modes. In particular, the short-term fluctuations associated with colloidal or hydrodynamic instabilities appear to be mainly random. They may be able to create conditions that favor the reappearance of similar pattern for some time, but there is no evidence that they can perpetuate themselves. This assumption—plausible but unproved—is basic to all existing theories of the general circulation or of climatic change.

Feedback effects might cause oscillations of the earth's surface-atmosphere-ocean system. Examples are the oscillations of ice caps discussed by Brooks (1949), or the variations of CO_2 content studied by Plass (1956). The characteristic time scale of these changes is probably too long, however, for them to be initiated by any transient individual circulation pattern.

Our basic assumption is equivalent therefore to a belief that external changes or long-term feedback effects will cause changes in the frequency of various circulation modes and consequently in the mean regime. This, in turn, will effect such characteristic aspects of the climate as mean cloudiness or mean snow cover. On the other hand, the exceptional weather situations or the freak individual season are not believed to exert a significant, finite influence over a period of decades on the future mean circulation pattern.

I shall discuss the character of circulation changes that might be expected to result from some slight changes in the mean distribution of some external variables. The expected changes then will be matched against actually observed climatic changes.

Radiation Balance and Heat Transports

The earth's surface gains heat, and the atmosphere as a whole loses heat, by radiation. Balance is maintained by air currents that transport heat from the surface upward to regions where it can be lost by infrared radiation into space.

The atmosphere can reduce the direct radiational heat loss from the surface into space because of its absorbing components: chiefly water, carbon dioxide, and ozone. If the amount of these gases was larger, the greenhouse effect would be stronger; that means the surface would tend to be warmer, more heat would necessarily be transported upward by air current, and more radiant energy would be emitted from the atmosphere. The radiational change would be most important at frequencies not fully absorbed in a thin air layer.

* Contribution No. 1186 from Woods Hole Oceanographic Institution.

There is an important difference between the radiative effects of these 3 absorbing gases. Carbon dioxide and water vapor, with a constant or upward-decreasing mixing ratio, cool the atmosphere at all levels. On the other hand, the mixing ratio of ozone increases upward in the troposphere, and the presence of this gas tends to heat the air there. Radiational cooling by ozone becomes effective only above the 30-km. level. Carbon dioxide and water vapor have an over-all destabilizing effect. Ozone radiation tends to destabilize the air near the ground and above 30 km.; it increases vertical stability through most of the troposphere and lower stratosphere.

Sensible heat is transported upward in the atmosphere by a rising of warm air and the simultaneous sinking of cold air. This process increases the overall static stability of the atmosphere and lowers its center of gravity. Conditions will remain steady if the upward heat transport and the resultant decrease in potential energy are compensated by differential nonadiabatic heating and if the generation of kinetic energy is equal to the frictional dissipation. A horizontal—or more precisely, an isobaric—flux of sensible heat does not generate kinetic energy.

The presence of horizontal heating gradients allows potentially warm air from low latitutides to spread above cold air from high latitudes. This causes the over-all stable stratification of the atmosphere. It gives the isentropic surfaces—that is, the surfaces of equal potential temperature—an upward slope from warm to cold regions and permits the development of quasi-horizontal baroclinic eddies.

From the work of Charney (1947), Eady (1949), and others it is known that this baroclinic instability is favored by a small vertical stability and large horizontal temperature gradients: a state characterized by relatively closely-spaced and steeply-sloping surfaces. We should expect, in this case, frequent, fast-moving disturbances of relatively small horizontal extent. On the other hand, if the ratio of downward to meridional potential temperature gradients is large, the isentropic surfaces will be more parallel to the earth's surface. Small perturbations will be resisted because vertically displaced air masses will lack buoyancy. Air has to come then from a great horizontal distance, where the temperature is sufficiently different, before it can become involved in a developing perturbation. Zonal propagation would be more sluggish, and we might expect a stronger development of blocking highs, monsoon circulations, and similar standing circulations: a condition that may be denoted as increased continentality.

Let p_h denote the pressure at the local mean upper limit up to which latent and sensible heat is being transported by air currents from below. Though closely related, p_h is not identical with the pressure of the tropopause. Particularly in low latitudes, penetrating convection causes a downward transport of heat just below the tropopause, and this makes the level of zero enthalpy flux slightly lower than the actual tropopause.

The correlation coefficient $c\{p_h, T_0\}$ between p_h and the surface temperature T_0 was shown by Goody (1949) to be negative. Therefore:

$$c\{p_h, T_0\} < 0 \qquad c\{\pi_h^*, \tau_0^*\} > 0 \tag{1}$$

where the bar denotes a space average over the whole planet at a moment of time, π_h^* is the horizontal variance of p_h and τ_0^* that of T_0.

Goody has shown also that the level of zero enthalpy flux would be raised in middle and high latitudes by an over-all increase in the CO_2 mixing ratio. It would be somewhat lowered over the tropics by more O_3 in the lower stratosphere. This means that the mixing ratio of both gases is negatively correlated with π_h^*; but an increase in O_3 would tend to increase p_h, while an increase in CO_2 would tend to reduce it.

The CO_2 mixing ratio and the O_3 mixing ratio at the tropopause level are both positively correlated with the mean surface temperature T_0. Their correlation with the meridional surface temperature gradient and therefore with τ_0^* is negative. In the case of ozone, this is accentuated by the larger concentration of this gas in high latitudes; in the case of CO_2, the 14-μ absorption band tends to have a more effective shielding action at lower temperatures. The decrease of the over-all static stability, which must be associated with any increase in the CO_2 mixing ratio should therefore be concentrated mainly in extratropical latitudes, and this would favor the development of baroclinic disturbances there.

Compared to p_h the distribution of surface pressure p_0 may be assumed to be approximately uniform ($\pi_h^* \gg \pi_0^*$). The over-all generation of kinetic energy $\dot{K}$ in the troposphere is given approximately by:

$$\dot{K} \sim k\hat{\bar{H}}[l_n(\bar{p}_0/\bar{p}_h) + \pi_h^*/2\bar{p}_h^2] \qquad \textbf{(2)}$$

The constant k depends only on the surface area and upon constant thermodynamic properties of the air; H is the vertical flux density of sensible heat, with the circumflex denoting a weighted vertical average.

The flux density of sensible heat has not been determined directly from observation. An estimate could be obtained indirectly from Equation (**2**), because $\bar{p}_h$ and π_h^* are observable and the value of $\dot{K}$ can be assessed with some confidence from estimates of the energy dissipation near the earth's surface. This yields a value of $\dot{K} \sim 0.5 \times 10^{22} - 1.0 \times 10^{22}$ erg sec.$^{-1}$ and of $\hat{\bar{H}} \sim 2000 - 3000$ erg cm.$^{-2}$ sec.$^{-1}$.

The vertical transport of sensible heat in the troposphere is likely to increase with increasing surface temperature and decreasing stability. The correlation between $\hat{\bar{H}}$ and the solar irradiation of the surface or between $\bar{H}$ and the CO_2 mixing ratio should therefore be positive. The value of $\hat{\bar{H}}$ would be reduced by an increased ozone concentration in the lower stratosphere and the upper troposphere.

An increased solar irradiation of the earth's surface would tend not only to reduce the mean downward gradient of potential temperature but would also increase the mean meridional gradient. This should lead to an over-all increase in kinetic energy generation and particularly to a more active development of baroclinic disturbances. An increase in ozone would tend to have the opposite effect: as discussed above, the greater stability and reduced meridional temperature variance would favor standing circulations and other indices of continentality at the expense of moving cyclonic circulation in this case.

Finally, an increase of CO_2 would tend to increase H and reduce p_h but it would tend also to reduce π_h^*. From Equation **2** it may be seen that the first two factors would make $\acute{K}$ larger; the third one would make it smaller. The over-all effect is probably one of increased $\acute{K}$. The decreased stability in middle and high latitudes would favor the incidence of baroclinic disturbances there, and this tendency would be amplified further by the effect of water in the atmosphere.

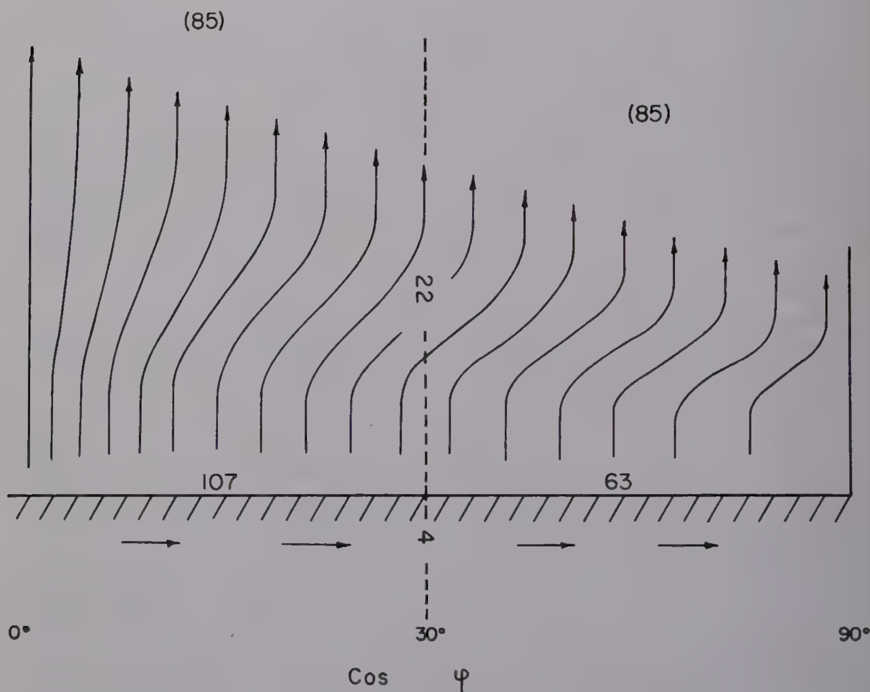


FIGURE 1. Schematic representation of heat transports in the northern hemisphere. Units are 10^{21} cal./yr.

The Circulation and the Hydrological Cycle

FIGURE 1 gives a schematic, order-of-magnitude representation of the latent and sensible heat transports in the troposphere and in the sea.

The net mean transport of heat across 30° N amounts to about 26×10^{21} cal./yr. Approximately 15 per cent of this transport, or 4×10^{21} cal./yr., is due to sea currents. The upward transport by air currents that originates at the surface amounts to about 340×10^{21} cal./yr. over the whole globe, or to about 10^5 erg cm.$^{-2}$ sec.$^{-1}$. Budyko (1956) states that about 82 per cent of this is latent heat, due to surface evaporation. The associated local heating of the atmosphere by condensation, as well as the radiative properties of the vapor itself and of the clouds, has a very large effect upon the distribution of heat sources and sinks, and therefore upon the circulation.

Both the CO_2 and the O_3 content of the atmosphere depend to some extent on factors that are independent of the circulation in the troposphere. The water vapor distribution on the other hand is so closely linked to atmospheric movements and temperature that it cannot be treated as an external influence. The amount of vapor that can be contained by the air decreases with decreasing temperature; it therefore decreases on the average upwards and polewards.

The radiational heat loss of the troposphere is closely related to the water vapor radiation from the layer immediately below p_h. The vapor pressure and temperature there cannot vary much with latitude. It is the mean temperature that would be registered by a celestial astronomer observing the earth in the 20 to 40 μ wave-length region. The total radiational heat loss through p_h appears to be more or less equally partitioned between the tropical and polar quarter of the troposphere, amounting to about 85×10^{21} cal./yr. for either, as indicated in FIGURE 1.

The radiational exchange between the atmosphere and the earth's surface also may be considered meridionally invariant as a first approximation. Although the surface radiates less energy at low temperatures near the poles, the back radiation from the atmosphere is reduced correspondingly because it originates from cold air layers that contain little water vapor. The net heat loss from the surface toward a cloudless sky may be estimated to be about 0.12 cal. cm.$^{-2}$ min.$^{-1}$ if the temperature is $^{-}10°$ C. with a typical surface air vapor pressure of 3 mb.; it has the same estimated value of 0.12 cal. cm.$^{-2}$ min.$^{-1}$ if the temperature is 25° C. with a typical vapor pressure of 16 mb. The variations in the intermediate range are also small. The mean, area-weighted, zonal, heat loss by infrared radiation varies as a result by less than 20 per cent from its global average (Budyko, 1956). The variations that do exist are mainly due to the meridional variation in cloudiness.

As the infrared heat exchange between the atmosphere and the underlying surface is relatively invariant, the meridional variation of the surface radiation balance must be associated mainly with changes in the irradiation. The total radiation that reaches any latitude zone depends on the angle of the sun and on the cloudiness.

The relation between cloud amount and the radiation balance varies not only with the latitude but also with day and night or winter and summer. The amount of radiation that reaches the ground at any given point and time may be considered an approximately linear-decreasing function of the relative cloud amount n. On the other hand, the effective outgoing radiation appears to decrease with n^m, where $m \sim 1.5 - 2.0$.

In low latitudes the incoming radiation is so much larger than the outgoing radiation that any increase in cloudiness would cause a reduced surface radiation balance there. It can be estimated that at the time of the equinoxes this would apply anywhere between about 52° N and S. At 60°, however, a cloud amount $n > 0.7$ would increase the surface radiation balance; at 70° any value $n > 0.4$ would have the same effect. It follows that an increase in cloudiness would tend to reduce the surface temperature in low latitudes and would tend to increase surface temperatures at high and mid-latitudes, particularly in winter.

The over-all cloud amount can be expected to be positively correlated with $\dot{K}$. This is due to the relationship between cloudiness and vertical motion that in turn is related to $\dot{K}$. From the discussion in the preceding section it may be deduced therefore that the cloudiness is likely to increase if the intensity of solar radiation reaching the earth's surface became stronger. The cloudiness increase must be small enough for excess surface irradiation to remain positive. If that were not the case, the increased convection and clouds could not be maintained.

An increase in the CO_2 mixing ratio was expected to result in decreased stability and more frequent baroclinic disturbances in middle and high latitudes. The associated increase in cloudiness would reduce further the radiational surface heat loss there. It would contribute therefore to a decrease in the mean meridional surface temperature gradient.

An increase of tropospheric and lower stratospheric ozone would tend to reduce the incidence of clouds. A reduced mean cloud amount is likely to accentuate the temperature difference between the oceans and the continents favoring further an increased continentality.

If much water vapor is transported upward by air currents the precipitation and evaporation also must be relatively large. The heat of evaporation is withdrawn mainly from the ocean. Increased evaporation therefore would lead to lower ocean surface temperatures, particularly in low latitudes and vice versa. An increase in ozone would tend to suppress or reduce the air currents that remove heat and moisture from the sea surface in low latitudes. It is therefore likely to cause a reduced meridional evaporation gradient. The available potential energy would be used at the same time for the maintenance of large standing circulations with fewer, weaker traveling, extratropical cyclones. This would favor the occurrence of meridional sea currents, such as the Gulf Stream, which provide an effective mechanism for the poleward transport of the excess heat from the subtropics.

An increase in CO_2 would tend to increase the meridional evaporation gradient, and the relatively greater frequency of baroclinic disturbances is less favorable for the maintenance of strong meridional sea currents. A larger CO_2 mixing ratio would therefore probably cause a larger fraction of the meridional heat exchange to occur in the atmosphere, and a smaller fraction to occur in the sea. Any increase in the intensity of solar radiation is likely to have the same effect.

Comparison of Derived and Observed Climatic Changes

The argument of the two preceding sections is summarized in TABLE 1, which gives the expected sign of the correlation between characteristic circulation variables on the one hand and, on the other, the hypothetical variations of the solar constant, the CO_2 mixing ratio, or the O_3 mixing ratio at tropopause level.

Variations of total solar energy emission in excess of 1 per cent have not been observed, and are improbable on astrophysical grounds, but this does not affect the present argument, which deals only with derived terrestrial effects. It should be noted that a hypothetical increase in the solar constant or in the CO_2 mixing ratio would lead not only to increased mean surface temperatures but also to

increased precipitation and decreased continentality. A reduction in the solar constant or in CO_2 content would of course have the opposite effect. Only changes initiated by ozone variations would cause increased surface temperature to be accompanied by reduced precipitation in low latitudes and vice versa.

The expected pattern of change, shown in TABLE 1, may now be compared with the actual pattern of change as derived from meteorological records over the past century. These actual changes have been summarized by, among others, Kraus (1958, 1960). They were characterized—up to about 1940—by an increase of approximately 0.01° C./yr. in the mean surface temperature. This increase appears to have been most pronounced in middle and high latitudes of the northern hemisphere, though it was not uniform in space or time even there. In low latitudes, precipitation decreased almost everywhere about the end of the last century. This decrease has received less attention than

TABLE 1

EXPECTED SIGN VALUE OF CORRELATION BETWEEN CHARACTERISTIC CIRCULATION VARIABLES AND EXTERNAL VARIABLES

Mean value of characteristic circulation variables	Solar constant	CO_2*	O_3†
Surface temperature	+	+	+
Poleward surface temperature gradient	+	−	−
Tropopause height	+	+	−
Poleward tropopause slope	+	−	−
Precipitation	+	+	−
Poleward precipitation gradient	+	−	−
Baroclinic instability	+	+	−
Continentality	−	−	+
Ratio of atmospheric to oceanic meridional heat exchange	+	+	−

* Mean mixing ratio.
† Mean mixing ratio in lower stratosphere.

the temperature increase. It is therefore documented by FIGURE 2, which has been reproduced from one of the above-cited papers. The ordinate in this diagram represents the function

$$y(n) = 100 \sum_{l=1881}^{n} \left(\frac{r_l}{\bar{r}} - 1\right) \tag{3}$$

where n is the running calendar year, l_1 the rainfall or stream discharge for the year l, and $\bar{r}$ the mean for the years 1881 to 1940. It is easily seen that y must be zero at the beginning of 1881 and at the end of 1940. The graph of y rises for $r_2 > \bar{r}$ and vice versa. It will be concave upward during the period when r is increasing with time, and convex when it is decreasing. The percentual deviation from $\bar{r}$ of the mean for any other period, for example, the years n to $(n + m)$, is given simply by $[y(n + m) - y(m)]/m$.

FIGURE 2 clearly indicates a higher precipitation level in the 19th century as compared with the first 40 years of the present century. The reduction of precipitation was associated with a reduced incidence of storms and a narrowing

of the tropical rainfall belt. It is best observed on the tropical boundary of the arid zone, but was less pronounced or absent in the monsoon regions. Conditions there and in the center of the big continents at higher latitudes were characterized by an increase of the January-to-July pressure amplitude, which

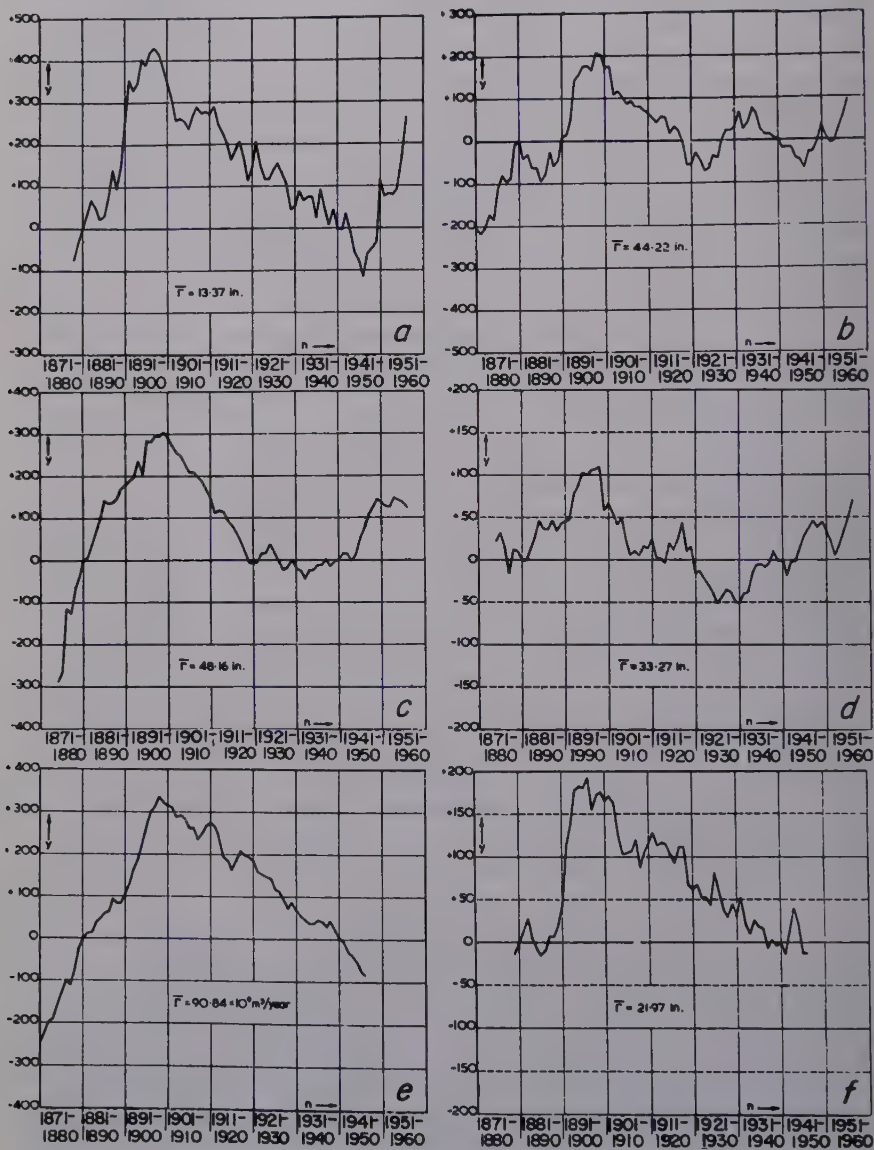


FIGURE 2. Cumulative percentual deviations from the 1881 to 1940 annual mean. Key: (*a*) N.W. New South Wales (Australian rainfall district No. 48); (*b*) Queensland Coast (Australian rainfall district No. 40); (*c*) Carolina Coast (mean of Charleston, S.C., and Cape Hatteras); (*d*) India (peninsula); (*e*) Nile river discharge at Aswan, Egypt; (*f*) Central Cape Province (South Africa rainfall district No. 16A).

suggests an increase in continentality. In the oceans, higher sea temperatures reported by Bjerknes (1959) and others were probably associated with decreased evaporative cooling in low latitudes.

During approximately the last 15 yr., the climatic trend appears to have been reversed. The warming in high latitudes seems to have ceased at least temporarily; at the same time most of the tropical regions and subtropical east coasts have become noticeably wetter again.

There is some evidence to suggest a Pleistocene glacial/pluvial sequence of basically the same character, although of larger amplitude than recent changes. Glacial ages were probably accompanied by wet conditions and therefore strong vertical heat flux in the tropics, and the end of the last ice age appears to have been associated with dryness at low latitudes and therefore with an increased meridional heat transport by sea currents.

If the arguments summarized in TABLE 1 are correct, it is not possible to explain either the pattern of recent climatic changes or the probable pluvial/glacial sequence of the Pleistocene by variations in the solar constant or in CO_2 content. Changes in the seasonal and meridional distribution of irradiation, caused by variations of the earth's orbital elements as calculated by Milankovitch, have not been considered here. They may have affected the Pleistocene sequence but they could not have produced the historic changes of recent times. The only type of variation that would be compatible with recent climatic changes—and that also could account qualitatively for the probable Pleistocene sequence—is the one initiated by ozone.

The quantity and the circulation of ozone in the upper stratosphere depends on solar ultraviolet radiation, which is known to vary to some extent, particularly at the shorter wave lengths. This suggests a possible link between climatic changes and the sun. However, any hypothesis would be on uncertain grounds until we know whether the existing ultraviolet variations can in fact influence significantly the formation of ozone and its transport to tropopause level.

Historical observations of the ozone distribution with height and latitude are insufficient for a test of the present argument. However, ozone is believed to have been possibly an agent of climatic change because of its effect upon tropospheric stability and tropical tropopause height. These last quantities have now been observed for a number of years, and this may permit a test of the present deductions with the aid of now available meteorological observations.

The changes that might be expected from CO_2 are different from those assumed to have occurred during the last million years. The possible result of a greatly increased CO_2 content is, however, of some speculative interest. It would be associated not only with a decreased meridional temperature gradient but also—because of high tropical evaporation and high polar rainfall—with an increased meridional salinity gradient near the ocean surface. This conceivably could lead to a reversal of the meridional circulation in the sea. Relatively warm, but highly saline and therefore dense, water would sink near the equator, to be replaced by fresher surface water from higher latitudes. This would be associated with a much higher temperature at depth, which would contribute additionally to a very high atmospheric CO_2 content. Conditions of this

type may have prevailed conceivably during a major part of the earth's history, when the bulk of the deep seas were known to have been in fact very much warmer than today.

Summary

The effect of external variations on the general circulation is discussed. Hypothetical variations of the solar constant would be positively correlated with mean surface temperature, mean meridional surface temperature gradient, mean tropopause height, and mean meridional slope of the tropopause. Variations in the CO_2 mixing ratio would be positively correlated with surface temperature and tropopause height; the correlation with the temperature gradient and the tropopause slope would be negative. The ozone mixing ratio at tropopause level also should be positively correlated with the surface temperature; correlation with temperature gradient, tropopause height and slope should be all negative.

The stability of the troposphere would be decreased by an increase in the solar constant or the CO_2 mixing ratio; it would be increased by additional ozone. Water vapor always tends to destabilize the air. Its presence requires additional upward transport of heat by air currents. Clouds reduce surface temperature in low latitudes and increase it in high latitudes. The correlation between the mean cloud amount and the surface temperature gradient is therefore likely to be negative.

Comparison shows that neither CO_2 nor solar constant variations are compatible with the pattern of recent climatic change, or with the probable pattern of the glacial/pluvial Pleistocene sequence. Variation of the O_3 mixing ratio in the troposphere and lower stratosphere would be compatible with the observed pattern, at least qualitatively. A large increase in CO_2 may be accompanied by a reversal of the meridional circulation in the sea, and this could explain the high temperatures of the deep oceans before the Pleistocene.

Acknowledgment

I am indebted to C. Rooth for discussion and helpful criticism.

References

BJERKNES, J. 1959. The atmosphere and sun in motion. : 65–73. Rockefeller Inst. Oxford Univ. Press, New York, N.Y. London, England.

BROOKS, C. E. P. 1949. Climate Through the Ages. McGraw-Hill. New York, N.Y.

BUDYKO, M. I. 1956. The heat balance of the earth's surface, Leningrad 255 pages (Translation prepared by Office of Technical Services, U.S. Dept. of Commerce). Washington, D.C.

CHARNEY, J. G. 1947. The dynamics of long waves in a baroclinic westerly current. J. of Meteorol. **4:** 135.

EADY, E. T. 1949. Long waves and cyclone waves. Tellus. **1:** 33.

GOODY, R. M. 1949. The thermal equilibrium at the tropopause and the temperature of the lower stratosphere. Proc. Roy. Soc. London. **A197:** 488.

KRAUS, E. B. 1958. Recent climatic changes. Nature. **181:** 666.

KRAUS, E. B. 1960. Synoptic and dynamic aspects of climatic change. Q. J. Research Meteorol. Soc. **86:** 1.

PLASS, G. N. 1956. The carbon dioxide theory of climatic change. Tellus. **8:** 140.

RECENT SECULAR CHANGES OF GLOBAL TEMPERATURE

J. Murray Mitchell, Jr.
Office of Climatology, United States Weather Bureau, Washington, D.C.

Introduction

Several years have elapsed since the most recent comprehensive studies of secular trends of global temperature were published. One such study of extraordinary interest is that of Willett (1950), who presented evidence of a net increase of world mean temperature that was especially rapid in the first decades of the 20th century and, apparently, still was in progress in the 1930s, the most recent decade of data available to him. In attempting to identify the ultimate causes of secular climatic variation, it should be ascertained whether this global warming trend has actually leveled off in recent years, as suggested by latter-day studies of Arctic data in the Scandinavian sector (Wallén and Ahlmann, 1955; Hesselberg and Johannessen, 1958) where the warming in earlier years was particularly noticeable. I consequently undertook to update Willett's study through the year 1959, using data for 1941 to 1950 in the recently published volume of *World Weather Records* (U.S. Department of Commerce, 1959) and global data since 1950 available at the National Weather Records Center. I took advantage of this opportunity to re-analyze Willett's basic data, after updating it, in order to compare the zonally-averaged secular trends in the Northern and Southern Hemispheres and in the tropics during the past century. The purpose of this paper is to summarize the results of this re-analysis of updated global temperature data and to make a preliminary evaluation of their significance to extant quantitative theories of secular climatic change.

Data and Method of Analysis

Following precisely the method of analysis adopted by Willett (1950), annual and winter* means of temperature since 1940 were averaged in consecutive 5-year periods (hereafter denoted as *pentads*) at each of a large number of stations distributed as uniformly as possible over the globe. The most recent pentad of Willett's study was that centered on 1937. The newly added pentads are those centered on 1942, 1947, 1952, and 1957. The new data were derived insofar as possible from the same stations selected by Willett for the 1937 pentad, supplemented by a number of more recently established stations that served to improve the geographical coverage of the overall global network, especially in oceanic areas. Still following Willett's procedure, the stations were then grouped by 10°-latitude bands. Within each such band, the differences of temperature between successive pentads were averaged over all stations. Cumulative sums of these average differences were then used as a measure of secular temperature variation in that latitude band. This

* The annual period is defined as beginning on January 1 in both hemispheres. Winter is defined in the usual manner as the months December through February (Northern Hemisphere) and June through August (Southern Hemisphere).

procedure minimizes the influence of secular changes of station network on the trend data. TABLE 1 lists the number of stations in each latitude band that contributed temperature-change data between successive pentads. Analogous data for earlier years were given by Willett (1950) in his Figures 2 and 3. The global and hemispheric average trends, shown in the next section, were derived as suitably weighted averages of the 10°-latitude band data.

Results

Global average changes. Let us first consider the trends of global mean annual and winter temperature, as shown in FIGURE 1. For this purpose the

TABLE 1

NUMBER OF STATIONS FOR WHICH CHANGES IN PENTADAL MEAN TEMPERATURE COULD BE COMPUTED, BY LATITUDE BANDS*

Latitude band	Pentads†			
	1937–1942	1942–1947	1947–1952	1952–1957
90–80° N	0	0	0	0
80–70	5	5	3	4
70–60	19	20	6‡	5‡
60–50	26	26	14‡	12‡
50–40	19	19	14	14
40–30	20	19	16	16
30–20	11	9	9	8
20–10	9	10	9	9
10–0° N	8	8	5	5
0–10° S	11	11	6	6
10–20	13	14	10	10
20–30	12	16	13	12
30–40	9	11	9	9
40–50	4	5	4	4
50–60	4	5	4	4
60–70	1	1	0	0
70–90° S	0	0	0	0
Totals	171	179	122	118

* Number in case of annual mean or winter mean data, whichever is less.
† Shows middle year of each pentad differenced.
‡ Soviet station data incomplete since about 1950.

latitude-band indices were averaged according to the geographical surface area of each band. In this figure, as in all following ones, the pentad centered on 1882 was recognized as the first pentad for which all latitude bands (except the polar extremities) were represented by some data. Consequently temperature changes are shown as departures from the temperature levels of the 1882 pentad. For comparison, Willett's data on global temperature change are shown in FIGURE 1 as dashed curves. It will be recalled that Willett's curves were derived by assigning equal weight to all stations regardless of the tendency for them to be concentrated in certain highly populated (northern) latitudes.

Comparison of the two sets of curves reveals several noteworthy facts. First, if the data are weighted by the area of each latitude band, the net global warming since the 1880s is found to be smaller in magnitude than Willett in-

dicated. Second, data since 1940, not originally available to Willett, show that the great warming of the earlier part of our century has not only leveled off but has apparently reversed to a substantial degree. The warming evidently culminated in the early 1940s, and global temperatures in very recent years have returned to the lower levels of the 1920s although they still exceed

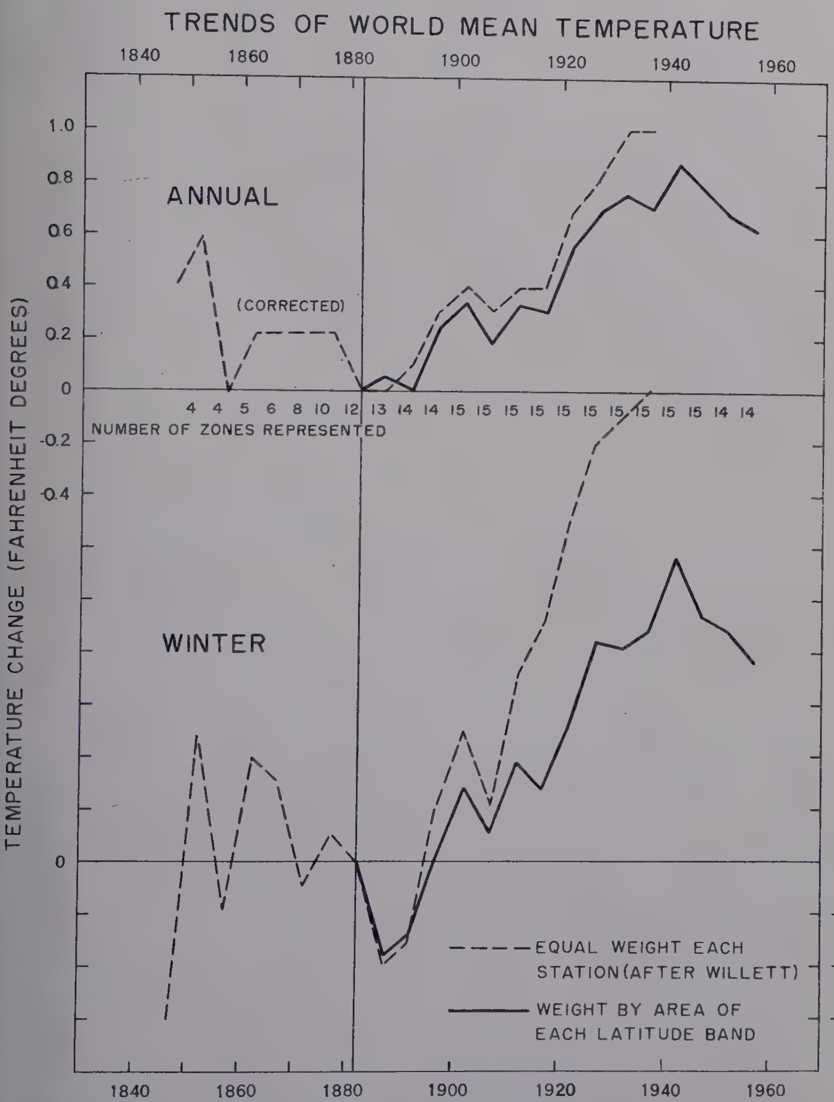


FIGURE 1. Trends of world mean temperature, shown as successive 5-year means relative to the 1880–1884 mean. Annual average data above, and winter-season data below. Solid curves are area-weighted averages of data for each of indicated total number of 10°-latitude bands, through the year 1959. Dashed curves are unweighted averages of data for all selected stations through the year 1939, after Willett (1950).

their long-term averages. Third, the rate of secular change of global average temperature, which Willett's form of the data showed to be uniquely large in the 1910-to-1930 period, is found from the present analysis to have been quite uniform throughout a much longer interval, namely between about 1890 and 1940. Quite obviously the question of whether the recent global trends are actually a segment of a long-period quasi-cyclical oscillation of global climate cannot yet be answered with confidence. Finally, it may be emphasized that the trends prior to the 1882 pentad, as determined by Willett, are based on a very meager and nonuniform geographical sampling of data and should therefore be regarded as speculative only.*

Comparison of Northern and Southern hemispheric changes. Next we may compare the annual and winter temperature trends in the Northern and Southern hemispheres. In view of the differing areas and distribution of ocean and land masses in the two hemispheres, and of the associated quantitative differences of their general circulations, it is not a priori obvious that the secular temperature trends of the two hemispheres should be equal. In comparing data for the two hemispheres, it must be realized that in the Northern Hemisphere data extend to much higher latitudes than in the Southern Hemisphere. Inasmuch as the trends of temperature are known to be a strong function of latitude (in the Northern Hemisphere at any rate), comparison of the two hemispheres should properly be based on equal segments of latitude in both. In FIGURE 2, average trends are therefore shown for 0–60° N, as well as for 0–80° N, along with those for 0–60° S. The secular trends of annual mean temperature for 0–60° N and 0–60° S are seen in the upper half of FIGURE 2 to have been remarkably similar since at least the 1880s. Shorter-period fluctuations in the two hemispheres, however, while they have been of comparable amplitude, have not always been synchronous. For the winter season, shown in the lower half of FIGURE 2, the situation appears to be quite different. In that season the Northern Hemisphere has experienced a much greater secular warming than comparable latitudes of the Southern Hemisphere, but the shorter-period fluctuations appear to have been similar in both phase and amplitude. Inclusion of data for 60 to 80° N in the Northern Hemispheric averages are seen in FIGURE 2 (dotted curves) to increase by about one half the net secular trend of that hemisphere, despite the small geographical area represented by the added data. In winter the excessive warming in the highest northern latitudes is seen to have set in first about 1920. In very recent years (since 1940), however, these same high latitudes have been the site of relatively rapid cooling (see also FIGURE 3).

Knowledge of conditions in Antarctica is thus far inadequate to establish whether temperatures in the highest southern latitudes have experienced a similarly exaggerated fluctuation in recent decades. Wexler (1959) has recently published data for the vicinity of Little America that, if representative of Antarctica as a whole, suggest a moderate net increase of mean annual temperature there of about 4° F. between 1912 and the present. It may be noted

* The change of annual temperature from the 1877 pentad to the 1882 pentad was plotted with a wrong sign in FIGURE 1 of Willett's paper (1950). The corresponding curve shown in FIGURE 1 here has been corrected for this error.

also that studies of temperature profiles in bore holes in the ice sheet near Mawson (MacRobertson Land) by Mellor (1960) are compatible with a secular warming of perhaps 2° F. per half century. While these results probably justify the conclusion that Antarctica has shared in the net global warming trend of the past century, they neither confirm nor preclude the possibility of

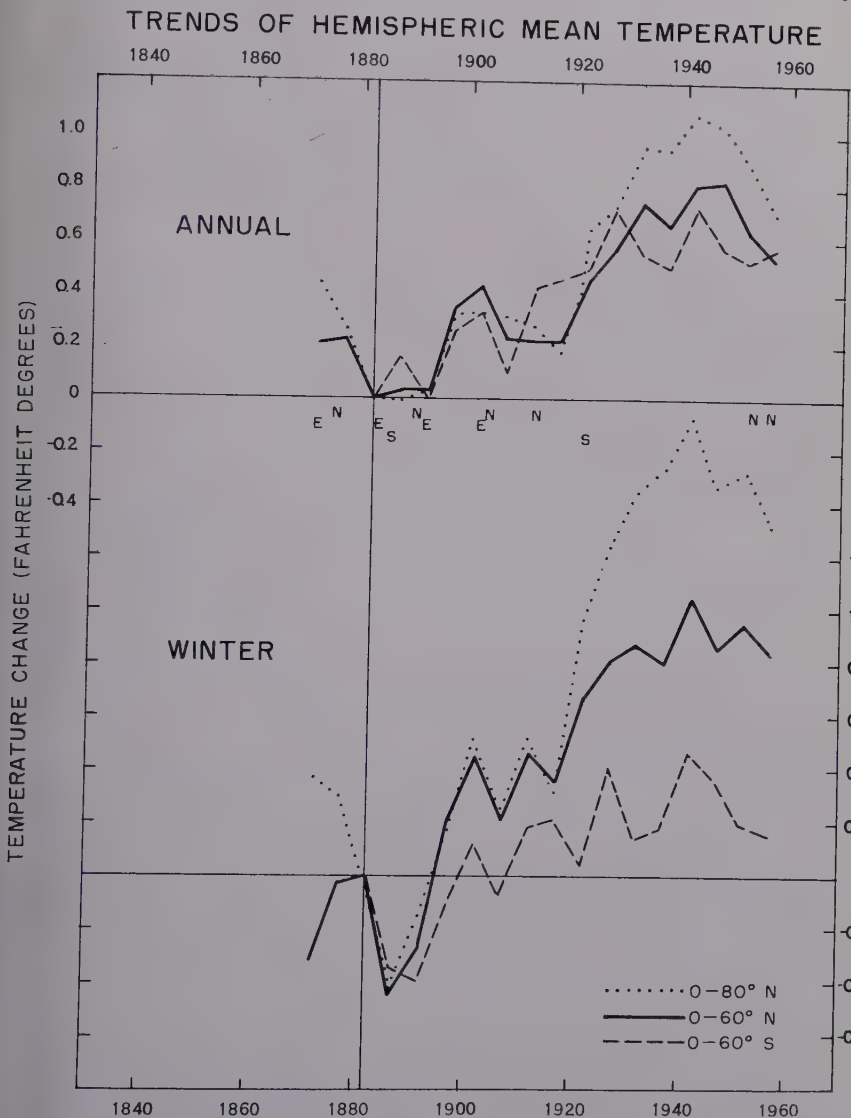


FIGURE 2. Trends of mean temperature in Northern and Southern Hemispheres, within indicated latitude limits. Annual average data above and winter-season data below. All curves are area-weighted averages of 10°-latitude band data. Dates of major volcanic eruptions shown in middle of the figure by E (for equatorial), N (for Northern Hemispheric), or S (for Southern Hemispheric).

larger-amplitude shorter-period fluctuations to compare with those observed in the Arctic.

Changes in the tropics. With so much attention focussed in recent years on the relatively large climatic fluctuations in the Arctic, it is well to point out that the tropics have also participated in the secular warming of the past

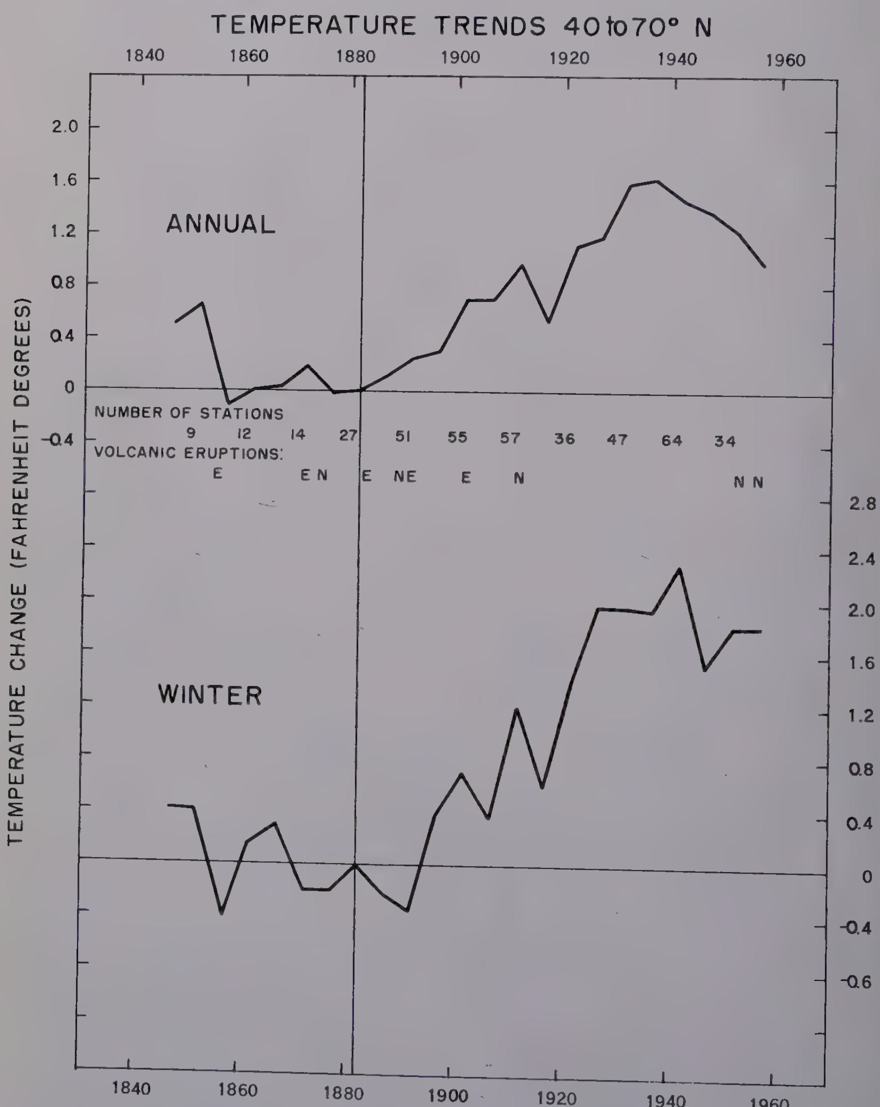


FIGURE 3. Trends of mean temperature in latitude zone 40 to 70° N, for which representative data (at the numbers of stations indicated) extend back to the 1845–1849 pentad. Dates of major volcanic eruptions in equatorial or Northern Hemispheric latitudes shown by symbols, as in FIGURE 2.

century. The trend of annual mean temperature in the zone between 30° N and 30° S, which represents half of the total surface of the earth, is shown in the upper part of FIGURE 4. Superposed on a net rising trend of about 0.8° F. since 1882 (the earliest pentad of fairly representative data coverage), shorter-

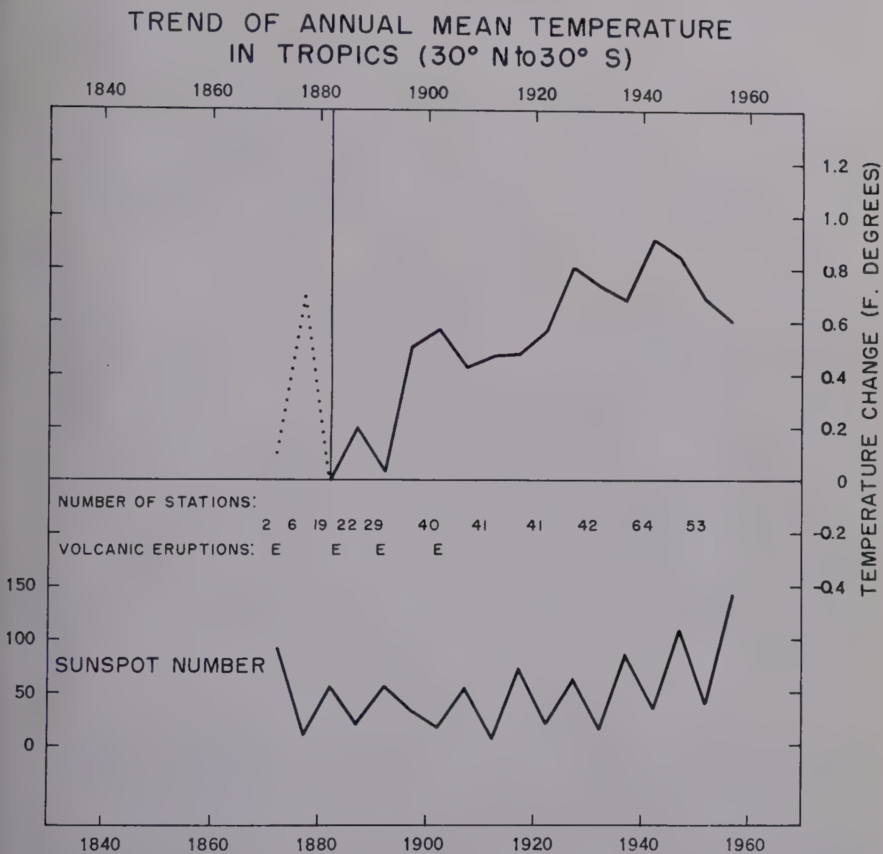


FIGURE 4. Trend of annual mean temperature in tropics (30° N to 30° S), the dates of major volcanic eruptions in equatorial latitudes (E), and successive 5-year averages of Zurich relative sunspot number. Temperature data based on numbers of stations indicated.

period fluctuations appear. We will remark further on these in the next section.

Significance to Theories of Climatic Change

Solar variability. From a comparison of the planetary blue magnitudes of Uranus and Neptune with those of carefully selected background stars, Johnson and Iriarte (1959) have pointed to evidence of a systematic increase of the solar constant from 1953 to 1958 totaling about 2 per cent. In the same period of time, Zurich sunspot numbers increased from a minimum, in early 1954, to

a record-high maximum in 1958.* From the point of view of this apparently direct relation between solar constant and sunspot number let us recall the record of sunspot numbers since the mid-19th century. First, the 11-year sunspot cycle implies that the solar constant has fluctuated with a similar period and phase of variation by perhaps 1 to 2 per cent.† Second, the secular trend of *average* (11-yr.) sunspot numbers, shown below in FIGURE 6, implies that the solar constant may have decreased by some 0.3 per cent from 1850 to 1900, and then have increased by about 0.7 per cent from 1900 to the present. In order to estimate the effect of such variations of solar constant on global mean temperature, we may apply the "broad-brush" method of Öpik (1953*a* and *b*), according to which the relative temperature variation $\delta T/T$ for a given latitude and season, resulting from a variation of solar constant $\delta I/I$, is given by

$$\frac{\delta T}{T} = \frac{1}{K}\frac{\delta I}{I}$$

where K, the "stability factor," is determined for various latitudes and seasons as a function of effective insolation; outgoing terrestrial insolation; the meridional transport of (sensible) heat due to the general atmospheric and oceanic circulations; the local storage of heat in ice, sea and soil; and surface temperature (see especially Öpik, 1953*b*). Using recent data on the Northern Hemispheric heat budget (London, 1957), and slightly different assumptions than those used by Öpik for the calculation of meridional heat transport from the surface temperature distribution, one finds annual values for K that range from about 4.5 in equatorial latitudes to about 15 in high latitudes of both hemispheres.‡ Thus the increment of hemispheric (or global) mean annual temperature associated with an increment ΔP of the solar constant (in per cent) is approximately§

$$\overline{\Delta T}\text{ (annual)} \simeq 0.8\,\Delta P \quad (^\circ\text{ F.}).$$

In order to estimate the corresponding effect on *winter* mean hemispheric temperature, a procedure was followed similar to that which Öpik followed in estimating the effect of solar-constant changes on summer mean temperature (Öpik, 1953*b*). By this means, much larger values of K are found to obtain in winter, and the increment of hemispheric mean winter temperature associated with an increment of the solar constant (in per cent) is found to be approximately‖

$$\overline{\Delta T}\text{ (winter)} \simeq 0.2\,\Delta P \quad (^\circ\text{ F.}).$$

* Based on data of Chernosky and Hagan (1958) and Lincoln (1959), the average sunspot number in the 12 months beginning November 1953 was 3 (minimum) and that in the 12 months beginning September 1957 was 203 (maximum).

† This conclusion is consistent with that of C. G. Abbott some 40 years ago, as noted by Ångström (1925).

‡ These may be compared with Öpik's original determinations of K, ranging from about 6 in low latitudes to about 20 in high. In the present application, this discrepancy is of minor significance.

§ This result agrees closely with an estimate by Fritz (1960).

‖ For comparison, the corresponding relation in summer is approximately

$$\overline{\Delta T}\text{ (summer)} \simeq 1.2\,\Delta P\ (^\circ\text{ F.}).$$

The conclusion that the thermal effect of solar-constant changes is greatest in summer and least in winter is a qualitative feature inherent in this type of analysis.

These results imply that the effect on global temperature of the secular variation of the solar constant, as inferred from the secular variation of sunspot numbers, is such as to account in the case of annual mean temperature for about a 1° F. variation in phase with the 11-year cycle, and a secular increase of about 0.7° F. since 1900. The corresponding thermal effects in the winter season would be parallel but much smaller in magnitude. Those in the summer season would again be parallel but larger.

How do the above conclusions square with observation? First, concerning the 11-year solar cycle, Köppen (1914) long ago cited evidence of a variation of temperature with the 11-year sunspot cycle, which attained its greatest amplitude of about 1° F. in tropical latitudes. However, Köppen concluded that this temperature effect is *inversely* related to sunspot number, which Ångström (1935) attributed to the overriding and opposite influence of variations of cloudiness that he believed also to follow the solar cycle.

Referring to FIGURE 4, in which 5-year averages of sunspot number are plotted along with those of annual mean temperature of the tropics, we find that Köppen's conclusion is amply verified in earlier decades but, curiously, not in the latter ones. This may be emphasized as follows. In order to minimize for the moment the complicating effect of secular trends in the two sets of data, let us consider the changes in temperature and sunspot number from each pentad to the next, and compare the signs of the concurrent changes in the two series. Of the 8 pairs of change values between the 1872 and 1912 pentads, *all 8 had opposite signs.* The probability of this relationship being spurious is only about 0.004.* However, of the 9 pairs of change values between the 1912 and the 1957 pentads, only 5 had opposite signs, which is not statistically significant. In any case, the average amplitude of temperature variation between pentads of high and low average sunspot number is seen in FIGURE 4 to be much less in either of the 2 epochs than the approximately 1° F. deduced by Köppen.

Concerning the secular trend of global temperature, it appears that a major fraction of the trend of annual mean temperature could be accounted for by the hypothesis of a secular variation of the solar constant following that of sunspot numbers. However, to the extent that Öpik's type of analysis is valid, the facts that the trends of temperature have been largest in winter, and largest at the higher latitudes, are not accounted for even qualitatively by such an hypothesis.

Carbon dioxide increase. It is by now quite reliably established that the average concentration of carbon dioxide in the atmosphere has secularly increased by somewhat more than 10 per cent since the 19th century (for example, Callendar, 1958), and that this increase is reasonably attributable to accumulation from the combustion of fossil fuel (Bolin and Eriksson, 1959). The consequences of such an increase to the heat balance of the atmosphere have recently been re-evaluated by Plass (1956), Kaplan (1960), and Kondratiev and Niilisk (1960). The quantitative results of each writer are quite different, depending in part on differences of assumed lapse rate and cloudiness and, most of all, on whether the joint effect of water-vapor absorption is taken

* It should, however, be realized that volcanic eruptions in tropical latitudes nearly coincided with three of the sunspot maxima in this period (see below).

into consideration. For the present purpose, I shall adopt the intermedial results of Kaplan, according to which the secular increase of CO_2-concentration derived by Callendar (1958) would result in a quasi-linear increase of global mean temperature totaling about 0.5° F. since the beginning of the 20th century. According to Kaplan's analysis, however, this increase should be relatively independent of season and latitude, in which respect some of the most important characteristics of recent secular changes are once again unaccounted for. On the other hand, Kondratiev and Niilisk (1960) concluded that the thermal effects of CO_2 changes should be largest where and when the water-vapor content of the atmosphere is small. While this conclusion might appear to bring the "CO_2 theory" of secular change into better qualitative agreement with observation, it should be emphasized that, in the view of Kondratiev and Niilisk, among others (for example, G. Yamamoto and S. Manabe, personal communication) the magnitude of influence of CO_2 variations on temperature is substantially smaller in the presence of any realistic atmospheric concentration of water vapor than the magnitude of influence derived by methods in which the role of water vapor is neglected (such as those of Plass and Kaplan).

Volcanic activity. Volcanic eruptions, at least those of catastrophic magnitude, have long been suspected of influencing large-scale climate (see especially Humphreys, 1920). By direct measurement, the dust injected into the stratosphere by such eruptions as those of Krakatoa (1883), Katmai (1912), and Kamchatka (1956) had depressed normal-incidence solar radiation by 10 per cent or more over intercontinental distances and for durations of several months. Although the influence of such eruptions on *total* (direct and diffuse) insolation is probably a little less than one half as much, the upward flux of terrestrial radiation is not significantly affected by the dust after a relatively short interval in which the larger particles have settled back to earth. Consequently significant widespread reactions of mean surface temperature to major eruptions are to be anticipated. Still in question, however, are the duration and magnitude of the temperature effect. Concerning first the matter of duration, the results of recent studies of the stratospheric residence time of radioactive debris following nuclear-bomb tests are particularly relevant. Stebbins (1960) has noted that effective half-residence times of such debris range from 5 months for the case of injections at polar latitudes, to between 10 and 20 months for the case of injections at tropical latitudes (depending on altitude reached). Inasmuch as the bomb debris probably penetrated to similar altitudes as the dust from the more violent known (explosive) eruptions such as that of Krakatoa, and as the small volcanic dust particles most effective in scattering sunlight, like the bomb debris particles, are so small as to possess negligible gravitational settling rates, we may infer that almost 90 per cent of the dust from such eruptions was removed from the atmosphere after 5 years. In the case of less violent equatorial eruptions, or of eruptions in higher latitudes, the same percentage depletion of dust probably occurred within 2 or 3 years.

The data summarized by Stebbins also indicate that debris injected into the stratosphere of one (the Northern) hemisphere disperses to virtually all latitudes of that hemisphere within a year, but only a very small quantity of it crosses the equator into the other (Southern) hemisphere. Consequently one may

conclude that an eruption in extratropical latitudes of one hemisphere does not spread dust into the other hemisphere in sufficient quantity to alter its heat budget materially.

On the basis of the foregoing conclusions, it is reasonable to expect that the global and hemispheric temperature data of FIGURES 1 and 2 should reveal something of the influence of major volcanic eruptions on temperature, despite the fact that the data are in the form of 5-year averages. In TABLE 2, known major eruptions since 1855 are listed along with the qualitative change of

TABLE 2

MAJOR VOLCANIC ERUPTIONS SINCE 1855 AND ENSUING CHANGES OF HEMISPHERIC MEAN TEMPERATURE

		Did world cool?	
		Annual	Winter
Tropical			
1855–6	*Cotopaxi, Ecuador†	yes?	yes?
1872	Merapi, Java†	no?	yes?
1883	Krakatoa, Java-Sumatra	yes	yes
1892	Awoe, Indonesia	yes	no
1902–3	Mont Pelée, Martinique and others	yes	yes
		Did eruption hemisphere cool? (Cool relative to other?)	
		Annual	Winter
Northern			
1875	Vatna Jökull, Iceland†	yes? (?)	no? (?)
1890	*Bogoslof, Aleutians	yes (yes)	yes (no)
1912	Katmai, Alaska	yes (yes)	yes (yes)
1953	Mt. Spurr, Alaska and *Hibok-Hibok Philippines	yes (yes)	yes (yes)
Southern			
1886	Tarawera, New Zealand	no (yes)	yes (yes)
1921	*Southern Andes	no (no)	yes (yes)
1932	Quizapú, Chile	yes (yes)	yes (yes)

* Denotes eruption of doubtful intensity.
† Data for Southern Hemisphere lacking.

temperature in the 5-year period following each eruption.* The results show a consistent tendency for eruptions to be followed by lower 5-year average temperatures in the eruption hemisphere. In order to estimate the average magnitude of this temperature effect, a superposed-epoch analysis was made in which the year of eruption was defined as the key date. For this purpose, the temperature changes of both hemispheres were given equal weight for the case of an equatorial eruption, but only the temperature changes of the eruption hemisphere were included for the case of an extratropical eruption. The

* In the event that an eruption occurred near the middle year of a pentad, the change of temperature associated with it has been estimated by considering the pentads preceding and following the eruption pentad, instead of two contiguous pentads. The reader is cautioned that some subjectivity was thereby applied in deciding which pentads to compare.

results of this analysis are shown separately for both annual and winter data in FIGURE 5. They indicate that, in terms of 5-year average temperatures, the quantitative effect of a typical major eruption—not necessarily explosive—is of the order of $-0.1°$ F., and that this effect is mostly if not entirely confined to the first 5 years following. These results are not inconsistent with the conclusion that planetary temperatures are depressed by as much as 0.5° F. or more in the first or second year following an unusually violent eruption and lend some credence, for example, to the possibility that the very low global temperatures of the latter 1880s and early 1890s were primarily attributable to the eruption of Krakatoa in 1883. It is also conceivable that the relatively rapid rate of cooling of the Northern Hemisphere within very recent years

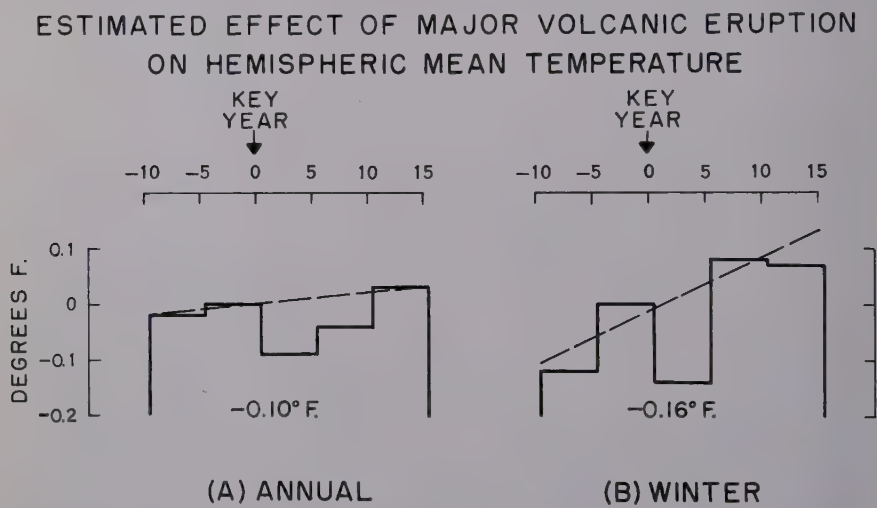


FIGURE 5. Estimated effect of major volcanic eruption on hemispheric mean temperature, based on superposed-epoch analysis of 5-year average temperatures shown as dotted and dashed curves in FIGURE 2, and by curves in FIGURE 3 for period prior to 1880. Key dates are eruptions listed in TABLE 2. Equatorial eruptions assumed to affect both hemispheres equally; other eruptions assumed to affect only eruption hemisphere.

(FIGURE 2) has been due in significant part to the Mt. Spurr, Alaska, eruption of 1953 and the great Bezymyannaya, Kamchatka eruption of 1956.* In brief the results of this analysis suggest that most of the transient irregularities in what otherwise appear to be highly regular, long-term variations of global mean temperature are attributable to the effects of volcanic activity.

Concluding Remarks

It is well to emphasize that the zonally integrated trends of temperature shown in this paper are not necessarily representative of particular geographical

* The vast extent of dust from the Mt. Spurr eruption was first suspected by Jacobs (1954) and by de Vaucouleurs (1954) party from evidence of anomalous illumination of the moon during the lunar eclipse of July 26, 1953. Subsequent examination of solar radiation records for 1953 by Fritz (1956) lent more direct confirmation of this. Concerning the intensity of the Kamchatka eruption, Gorshkov (1958) has ranked it with Krakatoa, Katmai, and Pelée. Dust from this eruption was visually observed over Great Britain (Bull and James, 1956).

regions. Owing primarily to contemporary trends in the position and intensity of the various cells of the general circulation, and to resulting changes in the temperature-advection field over the earth, secular temperature trends vary fully as much with longitude as they do with latitude (for example, refer to Figures 5 and 6 of Willett, 1950. This fact raises two fundamental questions:

First, is it possible that the foregoing evidence of variations of world mean temperature is merely the fortuitous result of *longitudinal migrations* of climatic zones over regions not equally well represented by climatological stations? Although the continents are much better represented by stations than the oceans, it appears that the Pacific Ocean is the only region in which the station network spacing is as great as the geographical scale of the secular change patterns. On the other hand, there is neither empirical evidence nor theoretical justification for supposing that the secular changes in the Pacific have ever been large enough to compensate entirely for the changes in other parts of the world for which data are more abundant. Consequently, if a completely uniform global network of stations were available whose data could be used in the present type of study, it is very unlikely that the *qualitative* results presented here would be vitiated. The magnitude of the global trends adduced from such a uniform network might, however, be somewhat smaller than that reported here.

Second, it is clear that the simple theories of secular climatic change thus far set down in quantitative fashion do not take adequate account of the feedback influence of changes of the general circulation. In this way, the apparent failure of solar variability and secular increases of atmospheric carbon dioxide concentration, to explain the observed dependence of secular temperature change on latitude and season, is by no means a valid reason for dismissing the possibility that such influences are fundamental causative factors in climatic change. For example, any factor that leads to an increase of the average effective heating of the atmosphere probably would also lead to increased evaporation over the world's oceans. This, in turn, would probably lead to increased cloudiness. In lower latitudes and in the warmer seasons of the year, this increased cloudiness would tend to offset the original temperature increase. In higher latitudes and in the colder seasons of the year, the increased cloudiness would tend instead to amplify the original temperature increase. The final effect on temperature would thus be in much better qualitative agreement with observations than the initial effect anticipated by the basic theories discussed in the previous section.

However, inasmuch as the theories of solar variability and carbon dioxide would each lead us to expect a continuing amelioration of world climate up to the present time, the cooling that has been in progress since the 1940s emerges as a curious enigma. The cause of this cooling, which cannot readily be traced to recent volcanic activity, remains for future research to clarify.

Summary

A 1950 study of world temperature trends by Willett has been updated through 1959, and the combined data are used to compare zonally-averaged secular variation in the Northern and Southern Hemispheres and in the tropics.

The warming of recent decades, especially marked in the early 1940s and in

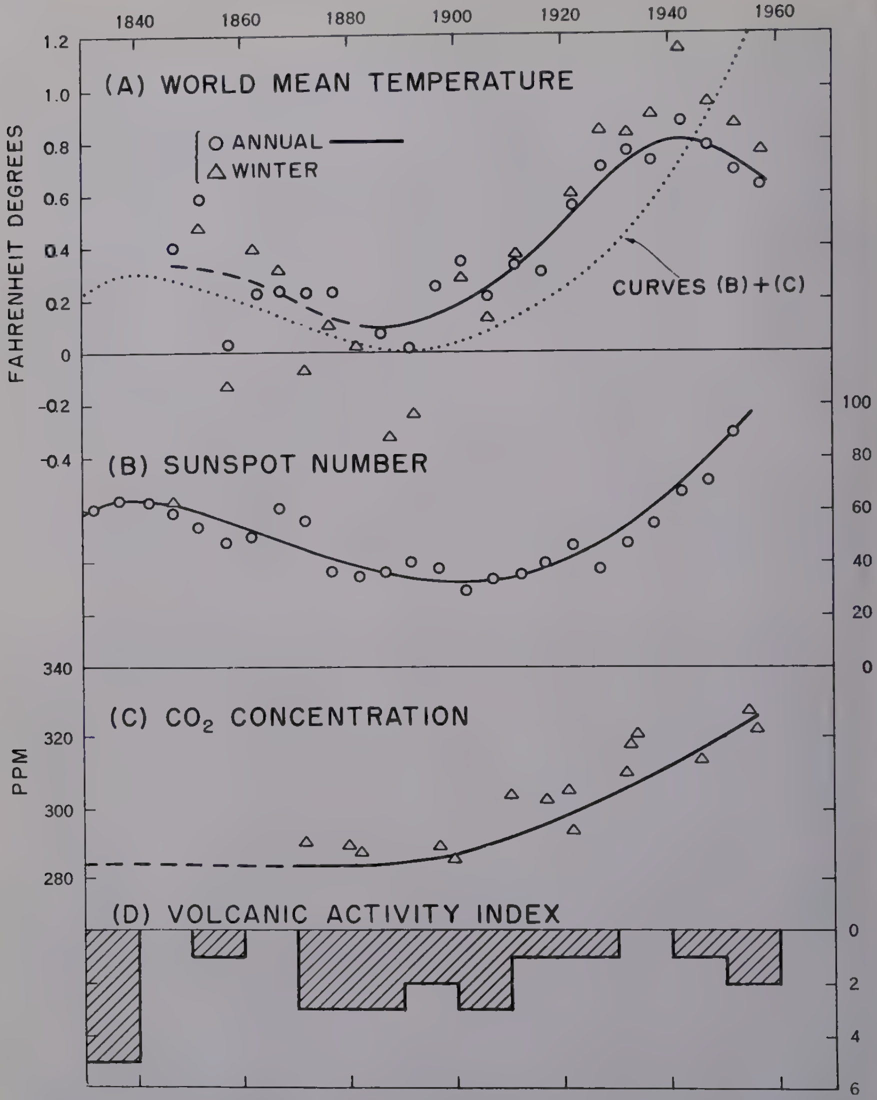


FIGURE 6. Secular changes of global mean temperature, sunspot number, carbon dioxide concentration, and volcanic activity during the past century. Sunspot data are overlapping 11-year averages of Zurich relative numbers. CO_2 data and fitted curve after Callendar (1958). Volcanic activity index after Lamb and Johnson (1959), inverted. Ordinate scales of curves *B*, *C*, and *D* adjusted to represent estimated magnitude of effect on annual mean temperature, according to ordinate scale of curve *A*. Scale adjustment of sunspot number based on solar-constant data of Johnson and Iriarte (1959) and thermal model of Öpik (1953*b*); that of CO_2 data based on calculations by Kaplan (1960), and that of volcanic activity based on empirical analysis in text.

winter, has since been replaced at most latitudes by moderate cooling. The magnitude of this cooling has thus far approached 30 per cent of the net warming from the 1880s to the early 1940s. Average global temperatures at present are comparable to those of the early 1920s.

Long-term trends are found to be parallel in the Northern and Southern Hemispheres, and transient depressions of temperature are found to be well correlated with the dates of major volcanic eruptions. It is estimated that hemispheric temperatures are depressed by the order of 0.1° F., averaged over the 5 years following major eruptions.

This paper concludes with some preliminary remarks on the significance of the observed global trends to extant quantitative theories of climatic change, including solar variability, the secular increase of carbon dioxide, and volcanic activity. Such theories appear to be insufficient to account for the recent cooling.

Acknowledgments

The global data on which this study is based were kindly supplied by H. C. Willett, Massachusetts Institute of Technology, Cambridge, Mass., for the period up to 1940, and by the able staff of the Weather Bureau National Weather Records Center for the period since 1940. I am indebted to both for their invaluable assistance. I also acknowledge the helpful guidance of Lester Machta, United States Weather Bureau, concerning the interpretation of recent thermonuclear fallout data and their value for estimating the stratospheric residence time of volcanic dust. Finally, I thank Ray E. Wilcox, United States Geological Survey, for his assistance in verifying the list of major volcanic eruptions shown above in TABLE 2.

References

ÅNGSTRÖM, A. 1925. On radiation and climate. Geograf. Ann. **7:** 122–142.

ÅNGSTRÖM, A. 1935. Teleconnections of climatic changes in present time. Geograf. Ann. **17:** 242–258.

BOLIN, B. & E. ERIKSSON. 1959. Changes in the carbon dioxide content of the atmosphere and sea due to fossil fuel combustion. *In* The Atmosphere and the Sea in Motion. : 130–142. B. Bolin, Ed. Rockefeller Inst./Oxford Univ. Press. New York, N. Y.

BULL, G. A. & D. G. JAMES. 1956. Dust in the stratosphere over western Britain on April 3 and 4, 1956. Meteorol. Mag. **85**(1012): 293–297.

CALLENDAR, G. S. 1958. On the amount of carbon dioxide in the atmosphere. Tellus. **10**(2): 243–248.

CHERNOSKY, E. J. & M. P. HAGAN. 1958. The Zurich sunspot number and its variations for 1700–1957. J. Geophys. Research. **63**(4): 775–788.

FRITZ, S. 1956. Opacity of the atmosphere after July 1953. Meteorol. Mag. **85**(1006): 110–112.

FRITZ, S. 1960. The heating distribution in the atmosphere and climatic change. *In* Dynamics of Climate. : 96–100. R. L. Pfeffer, Ed. Pergamon Press. New York, N.Y.

GORSHKOV, G. S. 1958. Neobychainoe izverzhenie na Kamchatke. (An unusual volcanic eruption in Kamchatka.) Priroda (Moscow). : 61–68.

HESSELBERG, T. & T. W. JOHANNESSEN. 1958. The recent variations of the climate at the Norwegian arctic stations. *In* Polar Atmosphere Symposium, Part I, Meteorology Section. : 18–29. R. C. Sutcliffe, Ed. Pergamon Press. New York, N.Y.

HUMPHREYS, W. J. 1920. Physics of the Air. McGraw-Hill. New York, N.Y.

JACOBS, L. 1954. Dust clouds in the stratosphere. Meteorol. Mag. **83**(982): 115–118.

JOHNSON, H. L. & B. IRIARTE. 1959. The sun as a variable star. Lowell Observatory Bull. No. 96. Flagstaff, Ariz.

Kaplan, L. D. 1960. The influence of carbon dioxide variations on the atmospheric heat balance. Tellus. **12**(2): 204–208.

Kondratiev, K. Y. & H. I. Niilisk. 1960. On the question of carbon dioxide heat radiation in the atmosphere. Geofis. Pura e Appl. **46:** 216–230.

Köppen, W. 1914. Lufttemperaturen, Sonnenflecke und Vulkanausbrüche. Meteorol. Z. **31:** 305–328.

Lamb, H. H. & A. I. Johnson. 1959. Climatic variation and observed changes in the general circulation. Geograf. Ann. **41:** 94–134.

Lincoln, J. V. 1959. Final relative sunspot-numbers for 1958. J. Geophys. Research. **64**(9): 1347–1349.

London, J. 1957. A study of the atmospheric heat balance. Final report, Contract No. AF 19(122)-165. Department of Meteorol. and Oceanography, New York University. New York, N.Y.

Mellor, M. 1960. Temperature gradients in the Antarctic ice sheet. J. Glaciol. **3**(28): 773–782.

Öpik, E. J. 1953. On the causes of palaeoclimatic variations and of the Ice Ages in particular. J. Glaciol. **2**(13): 213–218.

Öpik, E. J. 1953*b*. Convective transfer in the problem of climate. Geophys. Bulletin No. 8. School of Cosmic Physics. Dublin Inst. Advanced Studies. Dublin, Ireland.

Plass, G. N. 1956. Effect of carbon dioxide variations on climate. Am. J. Phys. **24:** 376–387.

Stebbins, A. K. 1960. A special report on the high altitude sampling program. Defense Atomic Support Agency. Washington, D.C.

United States Department of Commerce, Weather Bureau. 1959. World Weather Records 1941–1950. U. S. Government Printing Office. Washington, D.C.

de Vaucouleurs, G. 1954. Dust clouds in the stratosphere. Meteorol. Mag. **83**(988): 311–312.

Wallén, C. C. & H. W. Ahlmann. 1955. Some recent studies in Sweden on the present climatic fluctuation. Arch. Meteorol. Geophys. u. Bioklimatol. Ser. B. **6:** 7–21.

Wexler, H. 1959. A warming trend at Little America, Antarctica. Weather. **14**(6): 191–197.

Willett, H. C. 1950. Temperature trends of the past century. *In* Centenary Proceedings: 195–206. Royal Meteorological Society. London, England.

Addendum

After completion of the manuscript, a similar recent paper by Callendar (1961) came to my attention. Several essentially parallel results and conclusions are reached in both. Callendar understandably did not call attention, however, to the cooling in very recent years because his data ended with 1950, by which time the cooling had not yet attained noteworthy proportions. Further discussion of Callendar's interesting paper is in order, and is planned for a separate future note.

Reference

Callendar, G. S. 1961. Temperature fluctuations and trends over the earth. Quart. J. Roy. Meteorol. Soc. **87**(371): 1–12.

RECENT EVIDENCE ABOUT THE NATURE OF CLIMATE CHANGES AND ITS IMPLICATIONS*

I. I. Schell

Department of Geology, Tufts University, Medford, Mass.

Introduction

The solution of the problem of climate changes would be greatly aided by a better understanding of the nature of these changes, both during the earlier history of the earth as well as during the recent past which can be examined in greater detail. Although much remains to be accomplished before the true nature of these changes can be ascertained, the lengthening meteorological record, together with the other records from land and sea, makes it possible to probe deeper into the nature of these changes and into the validity of the different explanations that have been offered for them.

In seeking an explanation for a change in climate it was logical to consider that, since without the heat received from the sun the earth would be a cold body, this heat may have been "considerably diminished during glacial epochs" (Hitchcock, 1891). This very simple explanation of a change in climate poses a dilemma, as may be seen from the following quotation ". . . such a variation would be most effective in the low latitudes, but it is there that the climatic changes have been the smallest" (Brooks, 1950).

In explaining a climate change, such as, for example, an ice age, it becomes necessary, therefore, to reconcile the seeming paradox of a steeper equator-to-pole temperature gradient, increased circulation, evaporation, and precipitation with a decreased solar output that, because of the sphericity of the earth, seemingly calls for a greater decrease in temperature in the low than in the high latitudes and hence a less steep gradient, decreased circulation, evaporation, and precipitation.

In addition to a reconciliation between a greater decrease in temperature in the higher than in the lower latitudes and a decreased solar output, the explanation here offered must also account for (1) a change in climate that, although essentially global, is not simultaneous in all areas; (2) a more gradual onset of a cold period than of a mild period; (3) the long mild period preceding the Pleistocene and the transition to this epoch without, if possible, relying on extensive changes in the physical make-up of the globe; and (4) the smaller changes in climate or a climatic change.

As my first aim is to reconcile a decreased solar output with a greater lowering in temperature at the pole than at the equator, this point will be treated first.

A Reconciliation of a Greater Lowering of Temperature at the Pole than at the Equator with a Decrease in Solar Output

I shall try to show that while a slight increase in temperature gradient and in circulation will develop with a simple increase in solar output, a very much

* Tufts University Meteorological Studies No. 2.

The work described in this article was prepared with support from the National Science Foundation, Washington, D.C.

steeper temperature gradient and greatly increased circulation can occur with a relatively continuous decrease in the solar output. Three cases will be considered that for the purpose of this discussion may be treated simply.

Case 1: an increase in circulation with an increase in solar output. We remind ourselves that the effect of the sphericity of the earth is for it to receive more radiation from the sun at the equator than at the poles (see A–B, FIGURE 1), where a temperature of 26° C. is assumed at the equator and one of 0° C. at

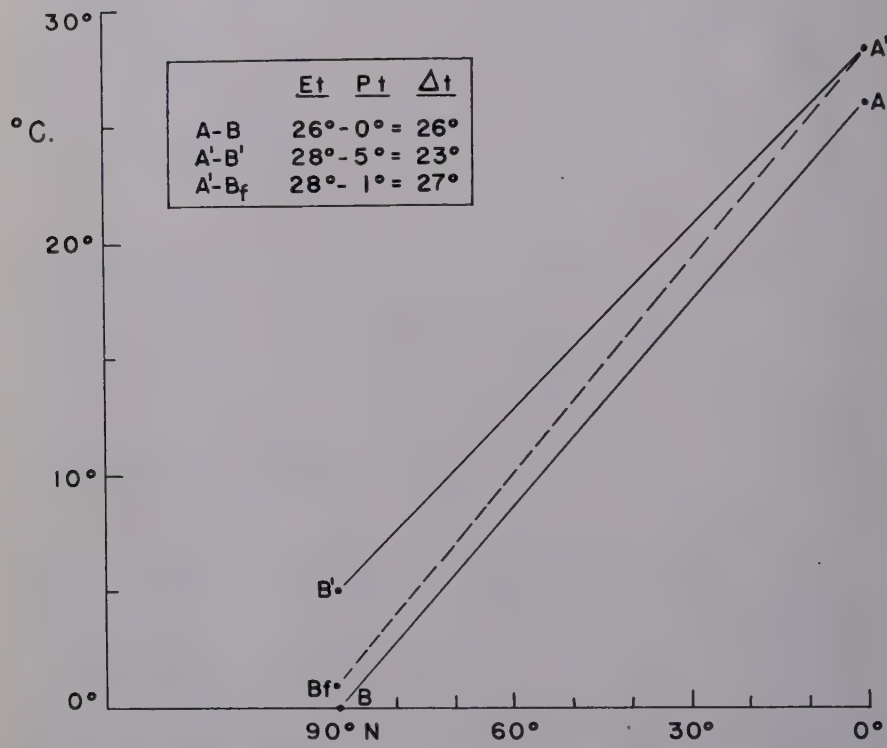


FIGURE 1. Equator-to-pole temperature difference following an increase in solar output. E_t, temperature of the equator; P_t, temperature at the pole; Δ_t, temperature difference.

the pole, roughly corresponding to the temperatures, respectively, during an interglacial (see below).

It is correctly assumed that if the solar output increased, the resulting increase in temperature, taken here as 2° C. at the equator would taper down from the equator to the pole, where the increase is, for convenience, shown as 1° C. (A^1–Bf, FIGURE 1). The steeper temperature gradient would lead to an increased circulation. However, following the increase in temperature at the equator there would first occur a larger increase at the pole, for convenience given here as 5° C. (A^1–B^1 in FIGURE 1), which is explained as follows:

As the poleward heat flux increases, it converges more and more over a smaller and smaller area, due to decrease of area with increasing latitude

(Defant, 1921). Therefore the immediate effect of an increased poleward heat flux would be a larger increase in temperature at the pole than at the equator, due allowance being made for divergence of the heat flux in transfer and for the increased heat lost by radiation at the pole due to its higher temperature.

The larger increase in temperature at the pole must now lead to a decreased equator-to-pole temperature gradient (A^1–B^1, FIGURE 1) and to a diminished heat flux. With no further increase in solar output, the temperature at the

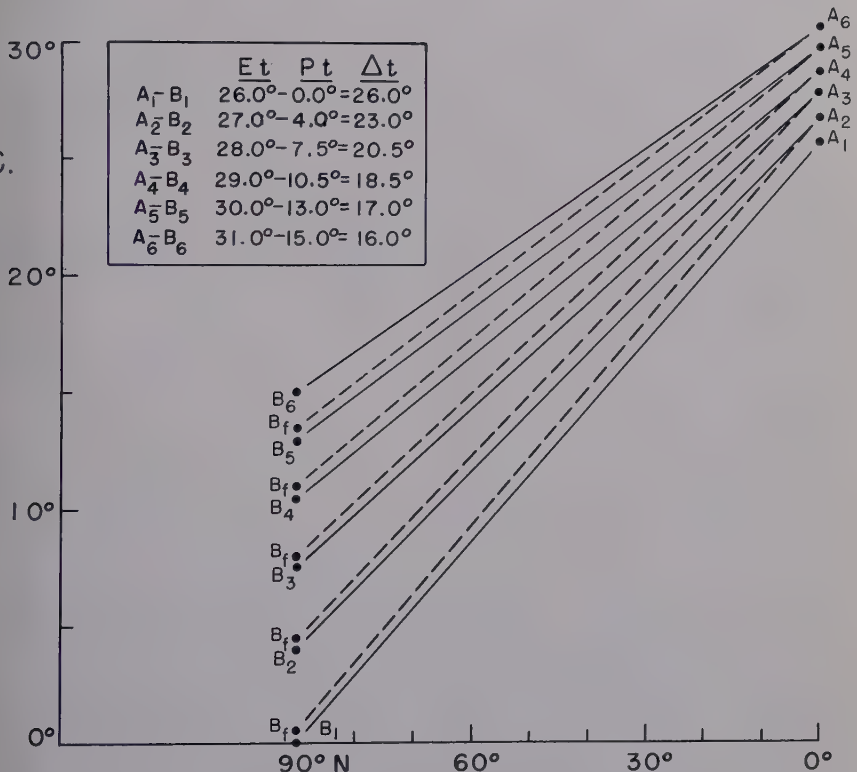


FIGURE 2. Equator-to-pole temperature differences following a relatively continuous increase in solar output. E_t, temperature at the equator; P_t, temperature at the pole; Δ_t, temperature difference.

pole would now drop to a slightly higher value than it was originally (A^1–B_f, FIGURE 1) in keeping with the initial increase in solar output. The final value of the temperature difference between the equator and the pole given here as 27° C. (FIGURE 1) would lead to a slightly increased circulation, evaporation, and precipitation, producing little change in the over-all climate.

Case 2: a decrease in circulation due to a greater increase in temperature in the high than in the low latitudes following a continuously increasing output. Now consider the same case as before, with a temperature of 26° C. at the equator and 0° C. at the pole (A_1–B_1 in FIGURE 2). Let us now assume an increase in the

solar output that is relatively continuous or such that occurs each time before the temperature at the pole can drop to a value consistent with the increase that preceded it. This will lead each time to a larger increase in temperature at the pole (B_2, B_3, $B_4 \cdots$, FIGURE 2) than at the equator (A_2, A_3, $A_4 \cdots$, FIGURE 2), the rate of the poleward heat flux becoming smaller and smaller as the equator-to-pole temperature gradient diminishes. Ultimately a condition will develop when the temperature at the pole is nearly the same as that at the equator and a tropical climate will prevail in the high latitudes. It must be emphasized that to obtain this result no large increase in solar output is required (see below) as long as the increases are relatively continuous.

Case 3: an increase in circulation due to a greater decrease in temperature in the high than in the low latitudes following a continuously decreasing solar output. Consider now the case of a relatively continuous decrease in solar output with a temperature 31° C. at the equator and 15° C. at the pole (A_6, B_6 in FIGURE 2). As a result of a relatively continuous decrease in solar output, the ensuing small drops in temperature at the equator and large drops at the pole (A_6 to A_5 to A_4, and downward and B_6 to B_5 to B_4 and downward respectively, FIGURE 2) will now lead to steeper temperature gradients (A_5–B_5, A_4–B_4, and downward, FIGURE 2) and to an increasing circulation.

Consider again a continuous decrease in the solar output with the temperature 26° C. at the equator and 0° C. at the pole (A–B, FIGURE 3). Again the temperature will decrease more at the pole than at the equator and a steeper gradient and increased circulation will develop (A^1–B^1, FIGURE 3). As the values of the temperature at the pole will now be below zero, a new factor is introduced: that of the ice which forms with below-zero temperatures. Because of the excellent radiative characteristics of this ice and the resulting cooling, the temperature at the pole will become even lower (B_i, FIGURE 3). As a result, the equator-to-pole temperature gradient will increase further (A^1–B_i, FIGURE 3), calling for a still greater increased circulation, evaporation, and precipitation. Also, with the over-all decrease in temperature and the increase in circulation and evaporation, the atmosphere will be now more heavily saturated and less able to hold its increased moisture. As a result, the moisture will readily precipitate, making room for additional moisture and continued precipitation.

If the solar output continues to decrease, it leads to other small decreases in the temperature at the equator and to large ones in the higher latitudes. However, with the outward spread of the ice, another factor is introduced: the shortening of the distance between the equator and the effective zone of the low (though not the lowest) temperature afforded by the edge of the ice. This decrease in distance between the zone of highest temperature at the equator and the zone of close-to-lowest temperature at some distance from the pole results in a further increase in the temperature gradient (A^1–B_d, FIGURE 3) as shown by the greater temperature difference over a smaller distance (29° C. + t_1 + t_2, FIGURE 3), and hence in a further increase in the circulation, evaporation, and precipitation. The increase in the precipitation "at the edge" of the ice would be enhanced by orographic lifting of the moist air, the ice in the process getting thicker while at the same time extending outward as the solar output continues to decrease.

These are the essential requirements for an ice age: lower temperatures at the equator and very much lower temperatures at the poles, a steeper equator-to-pole temperature gradient and a greatly increased circulation and evaporation, a lower level of saturation of the cooled air, and a greater precipitation

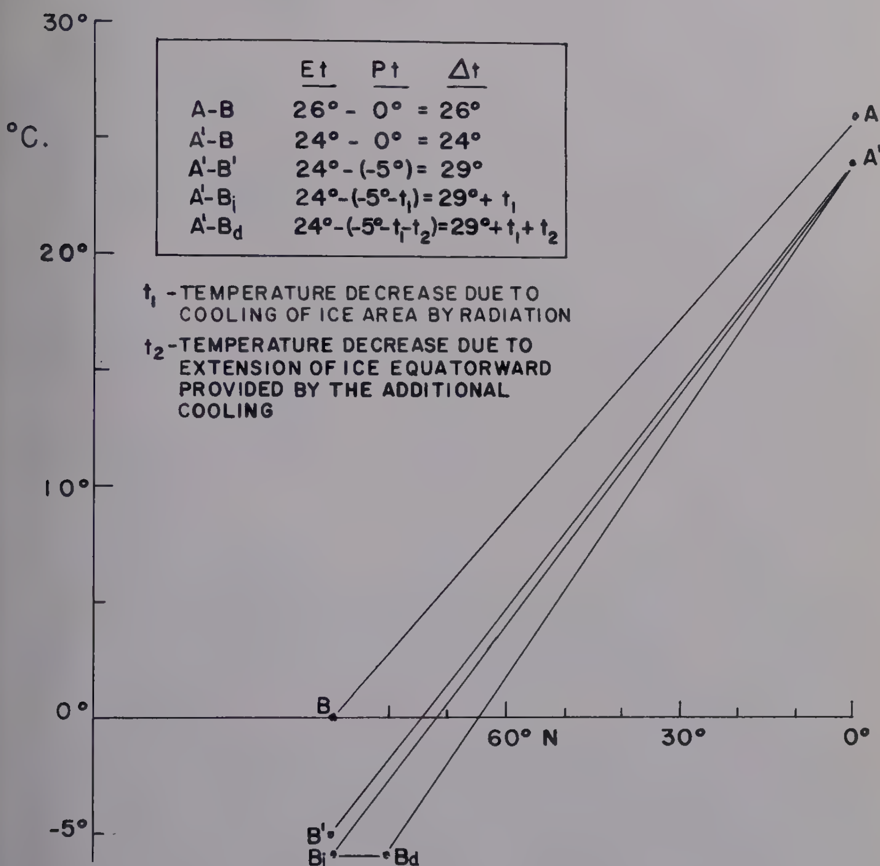


FIGURE 3. Equator-to-pole temperature difference following a relatively continuous decrease in solar output. E_t, temperature at the equator; P_t, temperature at the pole; Δ_t, temperature difference.

except over the ice areas as they expand and become dominated by anticyclones (see below).

Poleward Increase of Temperature Change

Introduction. As is well known, the temperatures in the higher latitudes are higher than the radiational temperatures, the actual temperatures being the result of the heat flux from the lower latitudes that because of the sphericity of the earth, converges poleward, leading to bigger temperature differences in them (see Defant, 1921). Similarly, with an increase in heat flux due to a rise

in temperature in the lower latitudes following an assumed increase in solar output, the rise in temperature in the higher latitudes would also increase poleward (Defant, 1921). Here I shall limit myself to showing that a small increase in temperature in the low latitudes is associated with a progressively bigger increase in the higher ones, leaving the question of linkage between changes in solar output and temperature, as in fact, variations in solar output itself, for consideration below.

Three cases will be considered: a short period increase in temperature from the relatively cool 1901-to-1920 period to the relatively mild 1921-to-1940 period; a long-period increase in temperature from a glacial to an interglacial in so far as it can be ascertained today; and the increase in temperature from the 1851-to-1900 to the 1901-to-1950 period for a more limited area.

Increase in temperature from the 1901-to-1920 to the 1920-to-1940 period. To show that a small change in temperature in the low latitudes is associated with a large change in the high latitudes, the temperature differences were computed between the first 20 years and second 20 years of this century, based on a network of stations, taken principally from *World Weather Records* (Clayton, 1927, 1934; Clayton and Clayton, 1947; Köppen and Geiger, 1939; Manley, 1953; Schuepp, 1958; United States Weather Bureau, 1959).

The stations chosen were carefully screened for completeness of record, continuity of site, hours of observation, and exposure of instruments, and where a significant change in any of the above had occurred, a suitable correction had been applied.

A total of 165 stations were thus available for comparison, of which 27 were in the Southern Hemisphere. Of the total number only 6 had a lower temperature in the 1921 to 1940 period, of which 4 were in the general Indian Ocean area, all with a decrease of 0.1° to 0.2° C., a fifth at Cape Pembroke in the Falklands, also with a decrease of 0.2° C., and finally the sixth, Laurie Island in the South Orkneys (Islas Orcadas), the most southernly station of all, with a decrease of 0.4° C.

TABLE 1, giving the average temperature difference between the two 20-year periods by zone of latitude, shows larger values in the higher latitudes (see also Lysgaard, 1949; Willett, 1950; Landsberg, 1960).

In the Southern Hemisphere, where the requirements of completeness and homogeneity of the temperature record limited study to fewer stations, the average increase from the first to the second 20-year period was not only smaller (TABLE 1) but, as already noted, a dip of 0.4° C. occurred in the South Orkneys (Islas Orcadas), as evidenced also by a marked increase in the ice severity in that area. Thus the Bay of Scotland was closed by ice an average of 213 days in the 1921-to-1940 period as compared with only 185 days in the preceding 1903-to-1920 period that marks the beginning of record (*Servicio Meteorologico Nacional*, 1951).

It is suggested that the smaller temperature increase in the Southern Hemisphere, sometimes known as the water hemisphere, is due in part to the greater amount of heat lost by downward transport over the oceans and to the large mass of ice in the Antarctic that would make the Antarctic and its surroundings less responsive to a relatively weak amelioration. Similarly the dip in

temperature in the South Atlantic polar sector would be due to the steeper temperature gradient and a somewhat intensified meridional circulation, as a result of the small rise in temperature in the lower latitudes, the circulation leading to greater outbreaks of cold air in that sector and to deeper intrusions of relatively warm air in the South Pacific polar sector, as indicated by a rise in temperature at Little America during recent decades (Wexler, 1959).

Increase in temperature from a glacial to an interglacial and equator-to-pole temperature differences. The temperature differences between a glacial and an interglacial in the low latitudes were quite small (Brooks, 1949) and, in the high latitudes, where the ice over many areas lay thousands of feet thick and later vanished, the temperature differences must have been enormous. Here an attempt will be made to indicate roughly what these differences were, using

TABLE 1
AVERAGE TEMPERATURE DIFFERENCE: 1921 TO 1940 MINUS 1901 TO 1920 BY ZONE OF LATITUDE

Zone	Temperature difference 0° C.	Number of stations
60°–80° N	1.43	18
40°–60°	0.43	65
20°–40°	0.31	40
0°–20°	0.27	15
0°–20° S	0.24	10
20°–40°	0.24	12
40°–60°	0.10	4
60°–	−0.40	1

A very substantial increase in temperature must have occurred also in zone 80° to 90° N as evidenced from the sharp decrease in the ice cover of the Arctic Ocean and reduced thickness of the ice, as well as from the increase in temperature of the Atlantic warm water beneath the cold layer under the ice during roughly the same period.

the results from ocean cores, change in height of the snow-line between a glacial and the "present," and other available evidence.

The increase in temperature from a glacial to an interglacial for western Europe in zone 40° to 60° N has been determined as approximately 11° C., a value composed of 8° C.—the difference between a glacial and the "present" and 3° C.—the difference between the "present" and an interglacial (Penck, 1938). The value 11° C. is probably close to the average for this zone, as over the North Pacific, which appears to have been little affected by glaciation, the difference must have been considerably less, while over the ice covered portion of North America in the same zone it must have been greater (see Manley, 1955), and greater also over the North Atlantic, as is to be inferred from a more southward extension of the ice over the ocean during a glacial.

Assuming the proportions of 8 and 3 between a glacial and the "present" and between the "present" and an interglacial, obtained for zone 40° to 60° N, to hold in the other zones as well, we may arrive at the temperatures in those zones, where in addition to the "present" temperature the upper or lower limit is also known.

The "present" temperature in zone 80° to 90° N is −19° C. (Hann and Suring, 1937), while during an interglacial when the Arctic Ocean was free of ice it was at least 0° C. (Brooks, 1949). Therefore the temperature difference between the "present" and an interglacial would be 19° C. Since this value would represent three elevenths of the total difference, the difference between a glacial and the "present" would come to 51° C., leading to a temperature of −70° C. during a glacial which, however, must have been higher due to the heat escaping through the ice from the water below (TABLE 2).

Similarly, the "present" temperature in zone 0° to 20° N is 26° C. (Hann and Suring, 1937). The average temperature difference between the glacials and interglacials of the Pleistocene came to 6° C., approximately, in the North Atlantic and to 1° C., approximately, in the eastern equatorial North Pacific (Emiliani, 1955). Additionally, the temperature differences computed from a change in the height of the snowline in equatorial Africa came to possibly as much as 5° C. (Flint, 1959) and, in the tropical Andes, to between 3° and 4° C. (Myer, 1904). Let us accordingly assume a difference of 4° C. for the zone as

TABLE 2

TEMPERATURE DISTRIBUTION DURING THE "PRESENT," A GLACIAL, AND AN INTERGLACIAL

	Present °C.	Glacial °C.	Interglacial °C.
80°–90° N	−19°	−70°*	0°
40°–60°	10°	2°	13°
0°–20°	26°	23°	27°

* The temperature over the Arctic Ocean was undeniably higher due to the escape of heat through the ice from the water below. Other temperature differences between a glacial and the present roughly in agreement with those indicated are: 7° C. at lat. 39° S in Australia (Browne, 1957) and 6° C. at lat. 37° N in Japan (Endo, 1933).

a whole, a difference admittedly open to revision yet consistent with the general result that the temperature in that zone during the Pleistocene changed little.

It follows that during a glacial with a mean annual temperature of −70° C. at the North Pole (actually higher) and about 23° C. at the equator (3° C. below the "present"), the equator-to-pole temperature difference would be approximately 93° C. and, during an interglacial, with a temperature of 0° C. at the pole and 27° C. at the equator (1° C. higher than the "present"), the equator-to-pole difference would come to about 27° C. or, roughly, one third of what it was during a glacial. Over the ice-free zone the temperature gradient actually would be much greater, since the temperature at the edge of the ice would be only a little higher than at the center of the ice cap, and the ice edge would extend many hundreds of miles equatorward.

Similarly, assuming for the South Pole the same difference of 70° C. between a glacial and interglacial and taking −49° C. (Flowers, 1958) as the "present" temperature, the temperature at the South Pole during a glacial would then come to −110° C. and, during an interglacial, to −30° C. and hence would be low enough for the ice in the Antarctic to maintain itself during the entire period of the Pleistocene (see Wright and Priestley, 1922; Gould, 1940).

The respective equator-to-pole temperature differences in the southern hemisphere would then be 133° C. and 57° C. approximately, substantially greater than the corresponding differences in the Northern Hemisphere.

Increase in temperature from the 1851-to-1900 to the 1901-to-1950 periods. Higher average temperatures were also obtained for the second 50-year period 1901 to 1950 than for the first 50-year period 1851 to 1900 for a small network of carefully screened stations in North America, Greenland, Iceland, Europe, and western Asia, as well as for 4 of the 5 stations in the Southern Hemisphere (Capetown, Buenos Aires, Sydney, Santiago), the exception being Adelaide, Australia, where a slight decrease, 0.1° C., had occurred in the latter period. The increase in the Northern Hemisphere ranged from 0.1° C. at San Diego and Tbilisi to 0.9° C. and higher at St. Paul, Toronto, Archangel, Stykkisholm, and Jacobshavn, the increase being smallest at stations with a marked maritime influence and/or lower latitudes (Bergen, 0.4° C.; Edinburgh and central England 0.3° C. each; also Prague, Zurich, Vienna, Bucharest, Sibiu, Kiev, Rome, Lisbon, Athens, Tbilisi, and San Diego with a range between 0.4° and 0.1° C.), indicating a poleward increase in temperature rise also during this period (see FIGURE 4). That the rise in temperature was not limited to the areas shown here is evidenced from the marked deglaciation in southern New Zealand, Equatorial East Africa, South America, as well as in southeastern Alaska, British Columbia, Europe, Iceland and Greenland during the same period (see Ahlmann, 1953).

Common Cause of a Change in Climate

By superimposing the temperature increase from the first 50-year period 1851 to 1900 to the second 50-year period 1901 to 1950 on a map of Pleistocene glaciation of the Northern Hemisphere (see FIGURE 4) it is evident that the larger temperature increases occur over areas where the ice on the whole was thickest and where, as one would expect, the contrast in temperature, after the ice has retreated, would be greatest. Similarly, the larger increases in temperature between the first 20-year period 1901 to 1920 and the second 20-year period 1921 to 1940 coincided on the whole with the areas in the higher latitudes occupied by ice during a glacial, the somewhat lesser agreement in the latter case being due to the increasing influence of the oceans and other factors, as the length of the period of change decreases (see below).

This suggests a common basic cause for long- and short-period changes in climate.

Relationships Between Changes in Solar Output and Climate

Introduction. The relationships between changes in the solar output and climate would obviously be more complicated than those considered for a simple earth above. They would be influenced among others by (1) the elevation of some of the areas which, for an equal decrease in solar output, would make the difference between a severe or light Ice Age, or merely a cool period; (2) the orientation of the continents and oceans, which permits a ready interchange of waters between some oceans and not others and the storage of heat in the surface layers during a warm period and increased stirring during a cold

period; (3) intense outgoing radiation from glaciated areas and a tendency for their outward growth during a cold period, and as follows.

Relationships with a decreasing solar output. Let us consider first the cli-

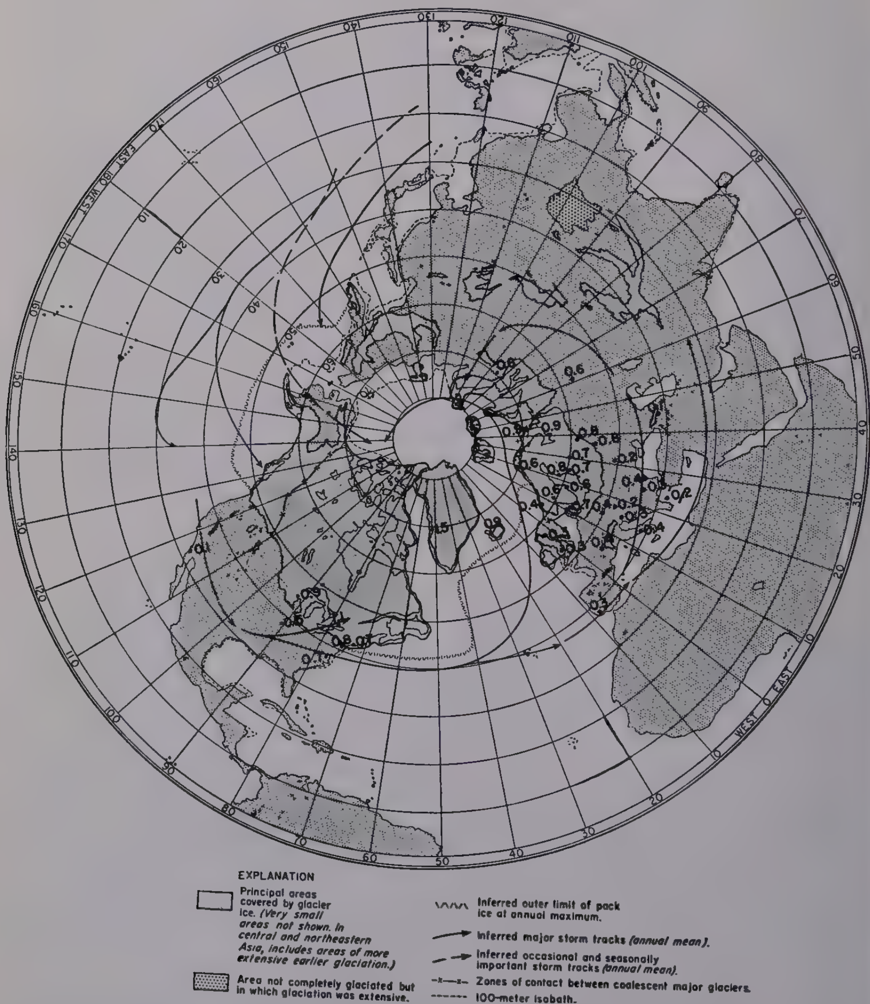


FIGURE 4. Average annual temperature differences: 1901 to 1950 minus 1851 to 1900 superimposed over map of Pleistocene glaciation. The values for the period 1851 to 1900 for Athens, Barnaul, Bucharest, Lisbon and San Diego, were based on less than the full 50 years. Reproduced by permission of John Wiley & Sons, Inc. (Flint, 1957).

mates associated with a decreasing solar output. As the Antarctic, because of its location and high elevation, would cool more rapidly than any other area, ice would form there first. The Arctic Ocean, because of its rather free exchange of water with the relatively warm Atlantic, would cool slowly, the ice making its appearance in that ocean well after it would have established itself

over the Antarctic. Should the solar output reverse itself and increase before the ice had a chance to form, the Arctic would then have experienced no glacial period, while ice would still be present over the Antarctic. Thus, because of: (1) the later establishment of ice in the Arctic, (2) the failure of the Arctic at times to reach the critical temperature necessary for ice formation, and (3) the earlier disappearance of the ice in the Arctic, with an increasing solar output (see below), the duration of the ice over the Arctic would be less than over the Antarctic for an equal period of time.

Furthermore, after the waters surrounding the Antarctic had been chilled they would become covered by ice for many miles outward. The very cold air flowing off the Antarctic and the adjacent ice-covered ocean would arrive still quite cool over southern South America, New Zealand, Tasmania and southernmost Australia and Africa, even after it has passed over some open sea; and during an extreme decrease in solar output the cool air would initiate moderate glaciation in the mountains of the areas lying closest to the Antarctic.

Ice would also be likely to form over Greenland and possibly over elevated portions of Scandinavia before it formed over the Arctic Ocean; similarly ice sheets would develop over the colder elevated areas in the middle and lower latitudes as, for example, the Alps, before the maximum advance of the ice from the north. From each of these areas in the Northern Hemisphere the ice would appear to creep northward to meet the ice advancing south, and in southern South America the ice would spread southward toward the ice advancing north across Drake Strait.

As a strong outflowing anticyclonic circulation would develop over the Antarctic, very little moist air would reach its "center." Then the ice at the "center" would be starved by evaporation, while the ice at the edge would continue to grow outward. As the Antarctic would be the first to develop an anticyclonic circulation, ice would be diminishing in its "center" while just beginning to develop over the Arctic and the elevated areas in high latitudes; similarly, the ice would be starving in the "center" of these areas while forming over smaller elevated areas in the low latitudes.

Hence, during prolonged periods of decreasing solar output, the ice sometimes would be forming over some areas after it has already formed over others; also sometimes first increasing and later decreasing in the same area. Bearing in mind also local topographic features and the varying stability of the ice, we should expect a complex pattern of glaciation, one that is in principle consistent with a decreasing solar output.

Again, because of the narrow opening between Siberia and Alaska, the ice from the Arctic Ocean would not readily extend into the North Pacific, and no sharp zone of contrasting temperature between the ice and the warmer ocean comparable with that over the North Atlantic, where the ice could readily advance southward, would develop. Hence the circulation over the North Pacific would be less vigorous, and relatively little moisture would be picked up from this ocean.

The principal source of moisture for the ice in the Northern Hemisphere during a glacial period would then be the North Atlantic, the moisture being carried westward and eastward by the easterlies and westerlies, respectively, associated with the North Atlantic low, the ice tapering off irregularly westward

over North America and eastward over Europe, also rising equatorward because of the richer source of moisture in the middle latitudes of the North Atlantic.

Also, partly because of the blocking produced by the Alaskan mountain ranges and, as already stated, because of the relatively small amount of moisture picked up from the North Pacific Ocean, little ice would form over Alaska, eastern Siberia, and adjacent areas.

The extensive precipitation associated with the storms just to the south of the ice edge would produce a tremendous run-off, which would fill the depressed land areas, create large lakes, and would be heaviest in the farthest zone reached by the ice, where the temperature contrast is greatest, or where the ice edge lingered longest. As the ice and the storm belt extended southward, the subtropical highs would be constricted and shifted equatorward, intensifying the tropical zone of convergence, while those over the smaller oceans, on reaching a critically small size, would be disrupted by storms. Thus over the larger oceans, where the high is in the main only constricted, there would be an intensification of the trades and an increase in upwelling in the equatorial zone, while over the smaller oceans, where the highs are disrupted by storms, there would be a decrease in upwelling, a much greater cooling by the cold air masses following in the wake of the disrupting storms, and a marked increase in precipitation. Over land in the equatorial latitudes, a less continuous and distinct convergence zone would develop, and the pattern of rainfall would be far more complicated. The small drop in temperature would lead to the formation of ice only over the mountains, the ice descending to the lower slopes.

In the areas between the storm belt away from the ice edge and the zone of tropical convergence, where the precipitation would be determined by storms irregularly breaking off the main track, a less distinct record of a pluvial age would result.

Relationships with an increasing solar output. Following a glacial period an increased solar output would lead to a small rise in temperature at the equator but to a progressively greater rise in the higher latitudes, leading to a deeper and deeper penetration of moister and warmer air poleward. Wherever the temperature of the air arriving at the "edge" of the ice would still be below zero it would cause an increase in snowfall; over the low-lying ice areas, where it would arrive at a temperature above zero, a decrease. If, however, the temperature continues to rise, the moisture would now everywhere condense more and more as rain. The ice caps and the anticyclones associated with them would then shrink until both had vanished. Thus the sequence in which the ice would disappear following a prolonged period of increasing solar output would be: (1) from the low-lying areas in the middle and high latitudes, as well as from the more elevated regions in the tropics; (2) from the Arctic Ocean and most of the elevated regions in the high and middle latitudes; (3) from Greenland; and (4) with a further and very prolonged period of increasing solar output and higher temperature, from the Antarctic itself.

Similarly, with the shrinking of the ice cap, the zone of sharp temperature contrasts between the ice, the open water, and the storm belt would shift poleward while the subtropical highs would greatly expand. This would lead to an increase in precipitation in the areas over which the storm belt is retreating

and to an increasing dryness over those now dominated by the expanded highs. As the ice disappeared the storm belt would vanish too, leaving only weak local disturbances where ocean and land, each characterized by somewhat different temperatures, meet. These small anomalies would then be the only departures from an otherwise zonal temperature distribution corresponding to a weak equator-to-pole temperature gradient, hence one associated with little circulation, advective evaporation, condensation, and precipitation.

With the ice gone, a further rise in temperature at the pole would from now on be determined solely by the increase in solar output rather than, as before, also by the intensely radiating ice. As the temperature at the poles approached that at the equator, the circulation would become sluggish and the atmosphere would approach a nearly quiet state. Conditions would then differ little from one place to another; convective clouds and heavy showers would be largely over the oceans and sunny skies for the most part over the land.

Also, the manner of exchange of air between the low and high latitudes would be relatively simple during mild periods, when neither ice caps nor the anticyclones associated with them would be present, and the temperature distribution nearly zonal. Then the sluggish exchange of air between the low and high latitudes would be little influenced by Coriolis force and the exchange would be more direct. During a glacial period the interchange of air between the low and high latitudes would be far more complicated and less direct due to (1) the presence of ice caps and permanent anticyclones over them, (2) areas of glaciation of various sizes in the lower latitudes, (3) constricted and greatly intensified anticyclones that would be disrupted over the smaller oceans by storms, (4) marked stirring in the oceans, and (5) a measurable Coriolis force.

Also, during a period of decreasing output and cooling, when the surface layers of the ocean would become too heavy to remain on top, the cold climates would take longer to establish themselves. It follows that for a period of increasing output of the same duration as a period of decreasing output the duration of the cold climate would be less than of the warm climate, since by the time the cold climate had established itself the output would have been decreasing for some time.

Similarly, the number of glacials or pluvials (very dry over the ice and very wet away from the ice) for a given number of equal decreases in solar output would not be the same in the equatorial zone as in the zone of shifting storm belt in the higher latitudes, where the change in climate between a glacial and interglacial would be more distinct; it would be also less in the areas between, where the increase in precipitation would be determined by haphazard storms. Thus the number of glacials or pluvials during a given epoch would vary from one broad zone to another, without in principle violating the number called for by the decreases in the solar output. Similarly, because of differences within the same zone, as between the North Pacific and North Atlantic, the intensities of the pluvials would also vary between these two oceans.

Comparison with the Facts

We may now compare the climates required by changes in solar output with the evidence gathered over many years about the glacials and interglacials of the Pleistocene (Charlesworth, 1957; Flint, 1957).

During a glacial the principal facts appear to be as follows:

(1) An apparent world-wide lowering of temperature, much more extreme in the high than in the low latitudes; the development of ice in the Northern Hemisphere, extending to latitude 38° N over North America, 52° N over Europe, 60° N over Asia; and the development of ice over the Alps and other mountainous areas in the middle and tropical latitudes away from the main ice sheet, as well as over southern South America and New Zealand, Tasmania, the mountainous areas of Southern Australia, and the islands in the vicinity of the Antarctic, and ice in the Antarctic that is thicker than today.

(2) The divergence of the ice from the elevated ice-covered areas in virtually all directions.

(3) The general tapering of the ice from east to west over North America and from west to east over Europe and Asia; its upward slope from northern Canada and Greenland and precipitous ending on the equatorial side.

(4) The abundant precipitation immediately south of the moving ice edge and the formation of large lakes in depressed areas, or their enlargement in both high and low latitudes.

(5) Little cooling of the North Pacific as a whole and active upwelling in its equatorial zone (Arrhenius, 1952); and markedly greater cooling of the North Atlantic and an apparent lack of upwelling in the equatorial zone, as may be inferred from a study by Emiliani (1955).

Similarly, during an interglacial the principal facts appear to be: (1) a world-wide rise in temperature, much greater in the high latitudes, with the ice disappearing from the Arctic Ocean (Brooks, 1949) and from virtually all other areas but persisting over the Antarctic (Wright and Priestley, 1922; Gould, 1940) and probably also over parts of Greenland; (2) a lower precipitation and drying up of many formerly wet areas; and (3) a world-wide and, in general synchronous, process of deglaciation, some areas experiencing a more rapid deglaciation than others, some experiencing an increase, most others a decrease.

The evidence further suggests a more rapid transition from a glacial to an interglacial than from an interglacial to a glacial (Büdel, 1949; Emiliani, 1955), in accordance with the demands for a greater retention of heat in the surface layers of the oceans with an increased solar output that would allow the quicker onset of a mild period.

Also such seemingly contradictory facts as the thinner ice cover, or absence of ice, in northernmost Greenland and northern Canada, the lack of any markedly greater precipitation in the Alps during a glacial than at present, as well as a possible lack of a decrease of ice in the Antarctic in recent years while the ice decreased elsewhere, would not be inconsistent with the demands of a decreased and an increased solar output, respectively.

The relative absence of ice in northernmost Greenland and Canada would be due to the fact that these areas would be starved of moisture as the ice edge advanced southward; they would become shut off from an inflow of moist air by the anticyclonic circulation that would develop over these areas while the existing ice cover was gradually reduced by evaporation.

Similarly, the apparent absence of greater precipitation over the Alps during a glacial than at present would be due to the development of an anticyclone

over the ice-covered area before the precipitation associated with the storm track south of the advancing ice edge over Scandinavia could markedly influence that area.

Finally the lack of a decrease in the ice in the Antarctic, or even a temporary slight increase in recent decades, if real,* would not be inconsistent with a general decrease in glaciation, because the precipitation from the increased inflow of moist air from the south would "at first" be in the form of snow as the air cooled on ascending this high barrier.

Relationships with Short-Period Changes in Solar Output

The relationships with short-period changes in solar output may be expected to be increasingly influenced by the oceans, elevation, glaciation, and the inertial forces inherent in such existing state of the atmosphere; these could conceivably overshadow the effect of the primary cause, or a short-period change in the solar output. Because a substantial portion of a change in the sun's output would be stored in the oceans, the contemporary relationship with the weather would be influenced also by the preceding solar output, severely limiting the relationship. Similarly, the contemporary relationship between the solar output and the weather would be affected by the internal relationships within the atmospheric circulation systems, as may be seen from the following.

With an increase in the southwesterlies over northwestern Europe, a rise in temperature would occur over that area that would at first be associated with a fall over East Greenland, due to the increased outflow of cold air from the Arctic over East Greenland in response to the greater inflow of air from the south. If the process continued, however, then the air flowing out from the Arctic would become less cold, leading to higher temperatures over East Greenland as well.

Therefore direct solar weather relationships would emerge chiefly with longer-period changes in solar output, as evidenced from the relationship between the moraines in southeastern Alaska and well-defined sunspot minima (Lawrence, 1951) and from the recent long-period rise in temperature with increased sunspot activity, here shown for the equatorial zone only (FIGURE 5), and assumed to reflect an increased solar output (see below).

From indications that the temperature differences between a glacial and an interglacial were very much smaller over the North Pacific than over the North Atlantic, it would appear that the effect of a change in the solar output, even of the order of tens of thousands of years, is not of the same intensity

* The evidence as to a possible change in the ice conditions in the Antarctic is very limited. Thus the shelf ice in King George VI Sound was observed to be 30 miles farther back in 1948 than in 1940 and still farther back than noted by the British Graham Land Expedition earlier in 1937; also, what was thought to be a single island (Eklund Island) in 1940 turned out to be one of 10 in 1948, the shape of the other 9 presumably becoming discernible as the ice level became lower during the intervening period (Fuchs, 1951). Similarly, the Northeast Glacier in places near Stonington Island has retreated 200 ft. in the interval from November 1940 to April 1947 (see Nichols, 1953) and, as recently reported, more than 1000 sq. mi. of Lady Newness Ice Shelf had disappeared since it was last charted in 1912 (Anonymous, 1959).

While citing lichens growing down to the edge of the ice on a nunatak inland in Queen Maud Land, as evidence of no decrease in the ice in recent times, Schytt (1953) concludes: "concerning the glacierization of Dronning Maud Land as a whole it can be said that no glacial retreat corresponding to that in northern latitudes is at present taking place."

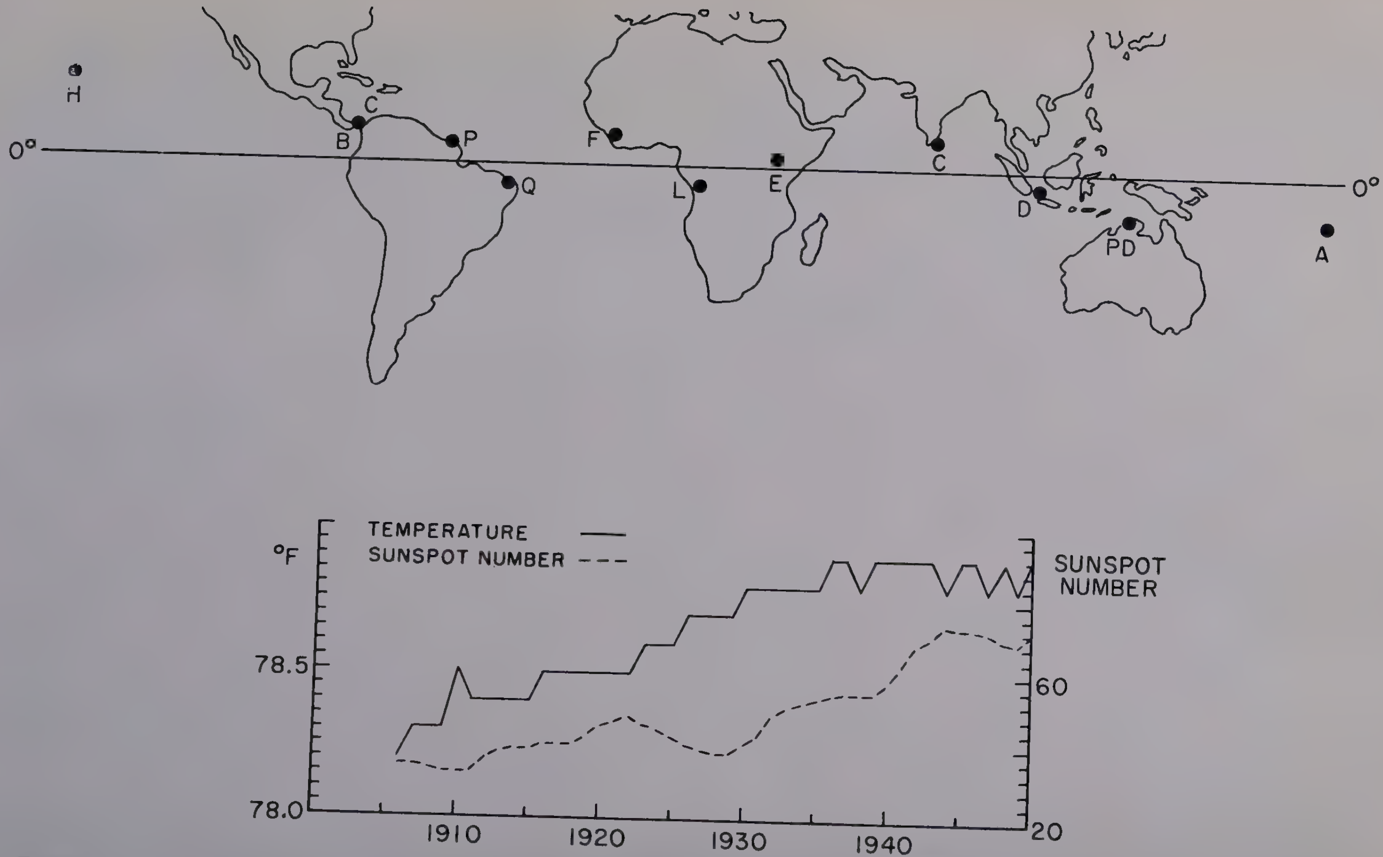


FIGURE 5. Annual values (overlapping sums of 11 years) of sunspot number and equatorial temperature (means of 11 stations: H, Honolulu; BC, Balboa-Cristobal; P, Paramaribo; Q, Quixeramobim; F, Freetown; L, Luanda; E, Entebbe; C, Colombo; D, Djakarta; PD, Port Darwin; A, Apia) during the period 1901 to 1954. As is well known, the observed rise in temperature in recent decades at sites in cities in middle and high latitudes was not primarily the result of their growth. At Djakarta, close to the equator at elevation 8 m., the rise in temperature from 1916 to

everywhere, the general rule being a certain nonsimilarity of all short-period weather patterns, due to the modifying influences of terrestrial factors and the inertial forces of the circulation, this nonsimilarity becoming less and less as the length of the fluctuation increases (see also Faegri, 1950; Kraus, 1955).

Variations in Solar Output and Temperature

The variations in solar output required to produce an estimated rise of 0.5° C. between the 1901-to-1920 and 1921-to-1940 periods, obtained by multiplying the temperature rise in each zone by the area and dividing the sum of the products by the total area, may be computed with the aid of the simplified formula shown below (Hoyle, 1949).

$$\frac{\Delta T}{Tm} \cong \frac{1}{4} \frac{\Delta H}{H}$$

where Tm is the mean temperature of the earth and ΔH the change in solar output, H, corresponding to the change in temperature, is ΔT.

Substituting in the above the present temperature 289° A for Tm, and the change in temperature, 0.5° C., between the two periods for ΔT, we obtain for ΔH approximately, 0.7 per cent, which is the net increase in solar output required to produce the rise measured between the two periods, one that is admittedly very small.

Similarly, for an estimated change in temperature of 11° C. between a glacial and an interglacial, the required change in solar output would be approximately 15 per cent over a period of many thousands of years.

Measurements of light reflected from the planets Neptune and Uranus show an appreciable increase in the solar output in recent years during which the sunspot number had similarly increased (Johnson and Iriarte, 1959) but the period is too short to allow a correlation.

Summary and Discussion

The evidence presented here indicates that the difference in temperature between a glacial and an interglacial in the low latitudes averaged not more than several degrees, and that between the shorter relatively cool and relatively mild periods it amounted to less than a degree, while in the high latitudes the respective differences were many times greater.

This fact is consistent with small but relatively continuous increases in the solar heat output and a poleward convergence of the heat flux due to the sphericity of the earth, for the loss of heat in transfer and loss by increased outgoing radiation in the high latitudes as a result of the higher temperatures is more than offset by the increased heat flux.

The evidence further indicates that the net increase in solar output required to produce a rise in temperature from a glacial to an interglacial was probably 15 per cent over a period of tens of thousands of years, and that the rise in temperature from the relatively cool to the relatively mild period in recent decades was probably less than 1 per cent, an amount so small as to make it difficult to detect with the present methods of measurement.

In regard to other possible influences on climate it may be pointed out that

the effect of continentality and elevation—in the sense that a glaciated area, for example, would chill its surroundings and thereby grow in size (Brooks, 1949)—is important, provided that the solar output remains constant. With an increase in output, the effect produced by the chilling of the glaciated area would be reversed and the ice would vanish. As we need not suppose that the solar output has remained unchanged for long periods of time, continentality and elevation by themselves could not have played a decisive role in climate changes unless the shape of the earth had been markedly different from what it is today. Furthermore, since an area of glaciation must be first initiated before it can grow outward, continentality and elevation cannot have been responsible for the initiation of an Ice Age; the solar output would determine whether the ice developed in the first place.

The pattern of temperature change from a glacial to an interglacial is like the change from a short cool to a short warm period whose basic characteristic is a change in the high latitudes many times greater than a change in the low latitudes. This pattern suggests that factors calling for (1) changes in the low latitudes opposite in sign to changes in the high latitudes, or for (2) changes in the northern hemisphere either unaccompanied by a change, or (3) a change different in sign from changes in the southern hemisphere, or for (4) changes of approximately the same magnitude everywhere, could only complicate somewhat the basic climate pattern called for by changes in the solar output. Indeed, the solar output itself might complicate the climate pattern far more than any of them.

Let the increase in the solar output, for example, be discontinuous, or let the time interval between increases be such that the temperature at the pole can drop to the level consistent with each increase before another occurred (see FIGURE 1), then the temperature at the pole will have risen less than at the equator, and the trend in climate will have been radically different—or perhaps even briefly reversed, although the net increase in solar output will have remained the same. The fact that a glacial and a mild climate could have developed during the long history of the earth suggests that the output of solar energy kept continuously increasing or decreasing, relatively, for long periods of time.

The climates that followed each other suggest that the sun's output was greater during the interglacials than the glacials, and still greater during the long mild period that preceded the Pleistocene; also that it gradually diminished during that period. The far greater occurrence of warm than of cold periods further suggests that large decreases in the sun's output were the exception.

The history of past climates also suggests that if the "present" is a true "present," a further warming is indicated, yet with every possibility of one or more severe coolings accompanied by an extension of the present ice cover and of other less severe coolings before the culmination of this interglacial. Since the temperature of the Antarctic appears to have remained well below the freezing point during the interglacials, except in its periphery (as may be inferred from a recent work by Pewé (1960)), we may think of this epoch of a single Ice Age with several major and many smaller ameliorations during which the ice disappeared each time everywhere except from the greater part of the

Antarctic and probably also from Greenland, and again, with many lesser ameliorations during which the polar ice caps and individual glaciated areas shrank in size in various degrees.

Summary

An analysis of the temperature changes between the 20-year period 1901 to 1920 and the 20-year period 1921 to 1940, the 50-year period 1851 to 1900 and the 50-year period 1901 to 1950, and a glacial and an interglacial of the Pleistocene, shows that the rise in temperature in each of the latter periods as compared with that preceding it on the whole increases poleward, the rise being consistent with a poleward convergence of an increased heat flux in the higher latitudes such as would arise from a small but relatively continuous increase in the total solar radiation or solar heat output.

It is implied that factors that call for a change in climate in the low latitudes, opposite in sign to a change in the high latitudes or for changes in the northern hemisphere, opposite in sign to a change or with no change in the southern hemisphere, or for changes of approximately the same magnitude essentially everywhere, could not account for the basic features of a climate change.

Acknowledgments

I am indebted to the directors of the meteorological services of Angola, Argentina, Australia, Brazil, Ceylon, Chile, Czechoslovakia, Great Britain, Hong Kong, India, Italy, Malagasy, Mauritius, Netherlands, New Zealand, Peru, Portugal, South Africa, Switzerland, Uruguay, and the West Indies for providing temperature data and their interpretation. I am also indebted to J. K. Charlesworth and R. F. Flint for reading this paper in part and for their valuable comments. I am also grateful to John Wiley and Sons for permission to reprint Plate 3 from *Glacial and Pleistocene Geology* by Richard Foster Flint, on which the temperature differences between the 1851-1900 and 1900-1950 periods were superimposed. This revised plate appears as FIGURE 4 in this report.

References

Ahlmann, H. W. 1953. Glacier variations and climatic fluctuations. Bowman Memorial Lecture Series. Am. Geog. Soc. New York, N.Y. **3:** 51.

Anonymous. 1959. Lady Newnes ice shelf. IGY Bull. **21:** 79–80.

Arrhenius, G. 1952. Sediment cores from the east Pacific. Swedish Deep-Sea Expedition 1947–1948 Reports. Göteborg, Elander. **5**(1)**:** 227.

Brooks, C. E. P. 1949. Climate Through the Ages. 2nd ed. McGraw-Hill. New York, N.Y.

Brooks, C. E. P. 1950. Climatic fluctuations and the circulation of the atmosphere. Weather. **5:** 113–119.

Browne, W. R. 1957. Pleistocene glaciation in the Commonwealth of Australia. J. Glaciol. **3:** 111–115.

Büdel, J. 1949. Die raumliche und zeitliche Gliederung des Eiszeitklimas. Naturwiss. **36:** 105–112, 133–139.

Charlesworth, J. R. 1957. The Quaternary Era. Arnold. London, England.

Clayton, H. H. 1927 & 1934. World Weather Records. Smiths. Misc. Coll. 79 and 90 (whole volumes). Smiths. Inst. Washington, D.C.

Clayton, H. H. & F. L. Clayton. 1947. World Weather Records. Smiths. Misc. Coll. 105 (whole volume). Smiths. Inst. Washington, D.C.

de Boer, H. J. & W. Ewe. 1949. On long-periodical temperature variations. Konin. Magn. et Meteor. Observ. Batavia. Verhandl. **35:** 1–16.
Defant, A. 1921. Die Zirkulation der Atmosphäre in den gemässigten Breiten der Erde. Geogr. Ann. **3:** 209–265.
Dirección General de Servicio Meteorologico Nacional. 1951. Datos Climatologicos y Geomagneticos. Periodo 1903–1950. Ser. B1ª Sec. No. 11. Republica Argentina.
Emiliani, D. 1955. Pleistocene temperatures. J. Geol. **63:** 538–578.
Endo, S. 1933. On the climate of the Pleistocene age in Japan. Am. J. Sci. **25:** 179–180.
Faegri, K. 1950. On the value of palaeoclimatological evidence. Centennial Proc. Roy. Meteorol. Soc. : 188–193.
Flint, R. F. 1957. Glacial and Pleistocene Geology. Wiley. New York, N.Y.
Flint, R. F. 1959. Pleistocene climates in eastern and southern Africa. Bull. Geol. Soc. Am. **70:** 343–374.
Flowers, E. 1958. Inside Antarctica No. 2—Amundsen-Scott-IGY Station. Weatherwise. **11:** 166–171.
Fuchs, V. E. 1951. Exploration in British Antarctica. Geograph. J. **117:** 399–421.
Gould, L. M. 1940. Glaciers of Antarctica. Am. Phil. Soc. Proc. **82:** 835–877.
Hann, J. von & R. Süring. 1937. Lehrbuch der Meteorologie. 5th ed. W. Keller. Leipzig, Germany.
Hitchcock, C. H. 1891. Discussion: The present standing of the several hypotheses of the cause of the glacial period. T. C. Chamberlin. Am. Geol. **8:** 237.
Hoyle, F. 1949. External sources of climatic variation. Quart. J. Roy. Meteorol. Soc. **75:** 161–163.
Johnson, H. L. & B. Iriarte. 1959. The sun as a variable star. **4**(96): 97–104. Lowell Observ. Flagstaff, Ariz.
Köppen, W. & R. Geiger. 1939. Handbuch d. Klimatologie. 3(N). G. Borntrager. Berlin, Germany.
Kraus, E. B. 1955. Secular changes of east coast rainfall regimes. Quart. J. Roy. Meteorol. Soc. **81:** 430–439.
Landsberg, H. E. 1960. Note on the recent climatic fluctuation in the United States. J. Geophys. Research. **65:** 1519–1525.
Lawrence, D. B. 1951. Glacier fluctuation in northwestern North America within the past six centuries. Assoc. Intern. D'Hydrol. Scientifique. UGGI. Brussels. : 161–166.
Lysgaard, L. 1949. Recent climatic fluctuations. Folia Geographica Danica. Kongelige Danske Geogr. Selskab. **5:** 215.
Manley, G. 1953. The mean temperature of Central England, 1698–1952. Quart. J. Roy. Meteorol. Soc. **79:** 242–261.
Manley, G. 1955. A climatological survey of the retreat of the Laurentide ice sheet. Am. J. Sci. **253:** 256–273.
Myer, H. 1904. Die Eiszeit in der Tropen. Geogr. Z. **10:** 593–600.
Nichols, R. L. 1953. Geomorphology of Marguerite Bay, Palmer Peninsula, Antarctica. Ronne Research Expedition. ONR Tech. Rept. **12:** 1–151.
Penck, A. 1938. Die strahlungstheorie und die Geologische Zeitrechnung. Z. Gessels. Erd. Berlin. : 321–350.
Pewé, T. L. 1960. Multiple glaciation in the McMurdo Sound region, Antarctica—a progress report. J. Geol. **68:** 489–551.
Schuepp, M. 1958. Der temperaturverlauf in der Schweiz seit dem Beginn der Meterologischen Beobachtungen. Ann. Schweiz Meteorol. Zentr. (Jahrgang 1957).
Schytt, V. 1953. Summary of glaciological work. J. Glaciol. **2:** 204–5.
United States Weather Bureau. 1959. World Weather Records 1941–1950. Washington, D.C.
Wexler, H. 1959. A warming trend at Little America, Antarctica. Weather. **14:** 191–197.
Willett, H. C. 1950. Temperature trends of the past century. Centennial Proc. Roy. Meteorol. Soc. : 195–206.
Wright, C. S. & R. E. Priestley. 1922. British (Terra Nova) Antarctic Expedition 1910–1913. Glaciology. **7:** 581. Harrison. London, England.

MAN'S ACTIVITY AS A FACTOR IN CLIMATIC CHANGE

H. Flohn

Deutscher Wetterdienst, Offenbach-Main, West Germany

Introduction

The question of the physical mechanism of climatic changes received increased attention about 30 years ago, when convincing evidence for recent climatic fluctuations was obtained. In the first decades of the present century the ice cover of the Arctic Ocean has decreased by more than 1 million km.[2] During the same period the glacier area of the eastern Alps has decreased by about one third, and the thickness of the glacier has shrunk about 60 cm. every year.

The problem of foreshadowing of climatic trends in the future is the most difficult task that can be put to a meteorologist. Since we know very little about the real physical causes of climatic fluctuations, little more can be done than to formulate a simple extrapolation of recent trends. There is little comfort in the thought that the meteorologists involved will hardly survive the success or failure of their forecasts.

Starting from the viewpoint of large-scale turbulent fluctuations -emphasized already by Defant (1921)—Manley (1953) has suggested a time spectrum of climatic variations, which will enlarge the time scale of the spectrum of meteorological phenomena toward larger values. If we consider the statistical relationship between time scale and area of meteorological phenomena—10^n sec. equivalent to 10^n or 10^{n+1}m—as extended into the region of climatic variations, we may ask whether the physical conditions responsible for climatic variations of longer duration—that is, 10^3 to 10^8 years—might extend beyond the limits of the earth. As a consequence of this, astrophysical hypotheses should be discussed as causes of such variations, but not necessarily of shorter fluctuations also.

During the 19th century, it was supposed sometimes that the apparent signs of climatic deterioration in arid and semiarid areas are caused by man's activity. After recent investigations of the thermal and water balance of the earth's surface it seems advisable to reconsider this hypothesis on the basis of our present knowledge.

Climatic Variations at a Constant Global Energy Balance

The first world-wide studies of climatic changes revealed the occurrence of large-scale variations in the strength and the position of the great action centers of the general atmospheric circulation, for example, of the subtropical anticyclonic cells (Wagner, 1940). From aerological evidence, the circulation patterns have been classified into at least two different types: high- and low-index types, according to Willett (1949). During World War II, a similar classification was used in Germany: zonal- and meridional-circulation types, having almost the same meaning. Recently Willett (1960) expanded this system by introducing a "low latitude zonal pattern." In fact the frequency of blocking anticyclones in the European section—a well-known example of low-index

type—has varied since 1881 in a 22- to 23-year cycle (Brezowsky *et al.*, 1951); these variations may coincide with the double sunspot or Hale period (FIGURE 1).

Of great interest is the variation of the resultant wind vector observed at two adjacent mountain observatories (Sonnblick and Zugspitze in West Germany), both near the 700 mb. level. Having corrected the records of Zugspitze according to local changes, a variation of the 700 mb. wind between the decades 1911 to 1920 (temperature maximum aloft) and 1938 to 1947 is found that can be interpreted (Flohn, 1951) as a shift of the quasi-stationary tropospheric ridge from a position near 20° E toward its present position near 5° E (FIGURES 2 and 3). This ridge is formed to a large extent by blocking anticyclones and may therefore vary with their frequency and positions.

Similar evidence has been demonstrated for the fluctuations of the positions of subtropical and subpolar belts (Wagner, 1940; Scherhag, 1936). These positions vary with the strength of the tropospheric westerlies which, in turn, is correlated with the meridional temperature gradient between subtropical

FIGURE 1. Annual frequency of blocking highs in the area between Iceland and Finland, overlapping 10-year averages. Data adapted from *Catalogue of European Großwetterlagen.*

and polar latitudes. The zonal index of the westerlies is also correlated with the number of tropical cyclones (FIGURE 4; Rodewald, 1957) and the date of onset of the Indian Summer Monsoon.

Two hypotheses, listed below, have been proposed as an explanation for these variations in the large-scale circulation patterns.

The feed-back mechanism of ocean atmosphere (Namias, 1959*a* and *b*). A regional anomaly of the ocean-surface temperatures, for example a cold-water pool in the Gulf Stream, leads to increased cyclogenesis in this highly baroclinic and frequently dynamically unstable region. As a consequence, we observe before these cyclones the building up of intense warm anticyclones in the eastern Atlantic and over Fennoscandia, and an increase of ocean currents in the North Atlantic, together with a large-scaled cold wave over North America, as well as other features over more distant areas. Berlage (1957) has suggested some apparently similar relationships in tropical oceans.

The influence of solar activity on terrestrial climate. This hypothesis cannot be dealt with in detail here (see Wexler, 1953; Willett, 1949).

These observed fluctuations of the general circulation (Lamb and Johnson, 1959; Kraus, 1960) can be interpreted as a redistribution of heat, precipitation,

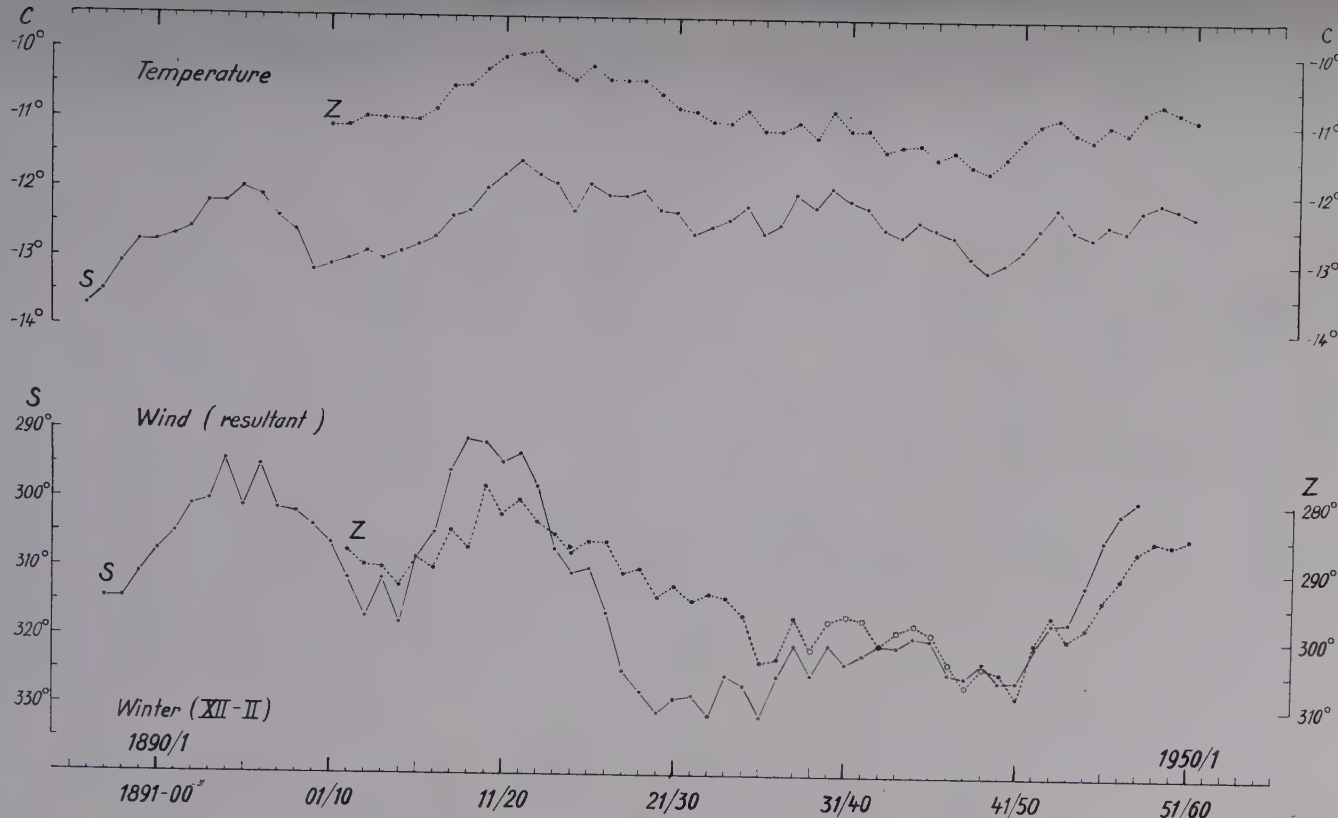


FIGURE 2. Average temperature and resultant wind at Zugspitze (*Z*) and Sonnblick (*S*) in the eastern Alps, overlapping 10-year averages for winter (December to February). Wind at Zugspitze reduced to the site since 1938 (blasting of the western summit) by a correction of −20° (*open circles*).

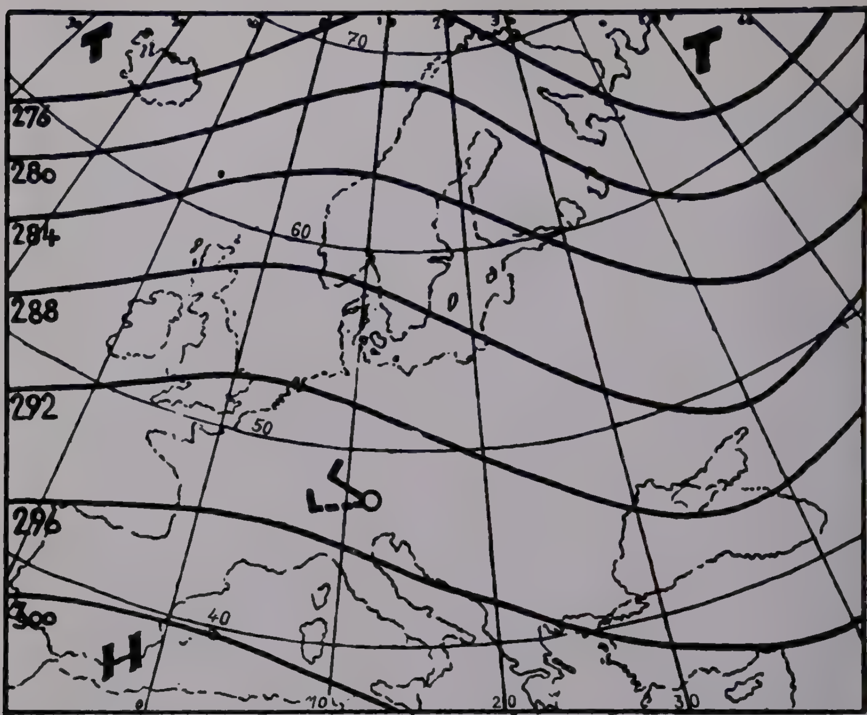


FIGURE 3. Average contours of the 700 mb. surface (unit = 10 geop. m.) in winter (December to February) and resultant wind at Zugspitze for the period 1939 to 1948; dashed-resultant wind at Zugspitze 1911 to 1920 (cf. FIGURE 2). Reproduced by permission of the German Meteorological Service.

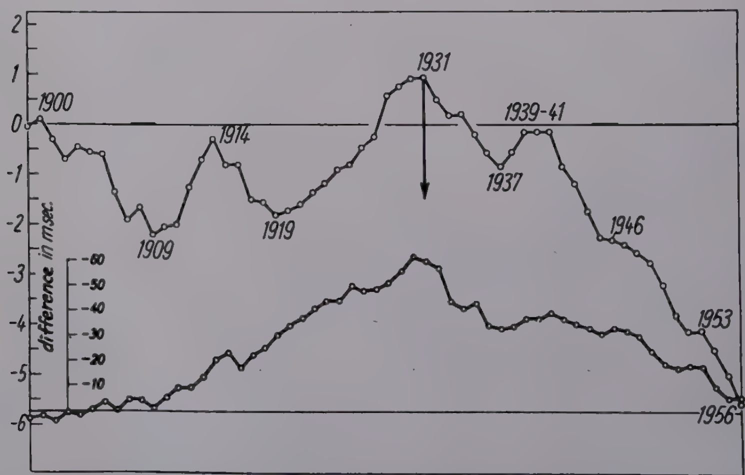


FIGURE 4. Frequency of hurricanes (*below*) and zonal index (35 to 55° N, 0 to 180° W), cumulative deviations from normal. Adapted from Rodewald (1957). Reproduced by permission of the German Meteorological Service.

and other such factors produced by quasi-stationary anomalies in periods where the global balance of energy can be considered as constant, at least from the practical point of view. The forecasting problem of those anomalous circulation patterns is in principle obviously the same as the problem of the long-range forecast. However, in recent years we have learned that the global energy budget is not completely balanced, and that we must envisage the occurrence of the storage of heat, wäter, and other such factors.

Climatic Changes and Fluctuations in the Energy Balance

As a matter of fact there is no convincing reason to assume that the energy balance of the climate-producing factors can be considered as constant, either on a regional or global basis. Therefore it has been suggested by the late C. G. Rossby that the geophysical energy budgets be investigated with special stress

TABLE 1

BALANCE EQUATIONS OF THE EARTH'S SURFACE

(1) Radiation balance:	Q	$= (S + H)(1 - a) - (E - G)$
(2) Thermal energy balance:	Q	$= U_B + U_N + U_L + U_V + F + \Delta U$
(2*a*):	ΔU	$= \Delta U_B + \Delta U_L + \Delta U_V$
(3) Water balance:	N_E	$= V_E + \Delta W$
(3*a*):	ΔW	$= \Delta W_M + \Delta W_K (+ \Delta W_L)$
(3*b*):	N_K	$= V_K + A_K \pm \Delta W_K$
(3*c*):	V_M	$= N_M + A_K \pm \Delta W_M$

Key: Q = radiation balance at the earth's surface; S = solar radiation; H = diffuse sky radiation; $(S + H)(1 - a)$ = effective global radiation; a = albedo of the earth's surface; E = terrestrial long-wave radiation; G = atmospheric long-wave counterradiation. $E - G$ = effective terrestrial radiation; U = (thermal) energy transfer; F = advective thermal flux (air)*; Δ = storage; N = precipitation; V = evapotranspiration; W = water, water vapor, or ice; A = runoff.

Indices: B = soil (or sea); L = air (sensible heat); V = evaporation (latent heat); N = warming of fallen precipitation + melting of snow and ice; E = earth; M = oceans; K = continents.

* On the use of the advection term F, see G. Hofmann (1960).

on the heat and water storage capacity of the atmosphere, the land, and the oceans. This suggestion provides us with a rational physical basis for a systematic discussion (Mitchell, 1953; Veryard, 1956) of theories of climatic change (Flohn, 1957, 1958), and it may protect us from overemphasizing one or another individual factor. As a consequence we are obliged to deal with the problem from its very roots, and we are confronted with our ignorance, at least regarding the numerical values of some basic properties, and with the inevitable need of much more representative and reliable data.

We may write down the energy balance equations of the atmosphere as shown in TABLE 1. In these basic equations only biological processes (photosynthesis and organic decay) are neglected. Starting here we may briefly consider individual terms in their possible effect on climatic changes.

(1) The hypothesis of a recent increase of the solar constant S_0 of the order of 0.01 per cent per year (Aldrich, 1950) has not met with general agreement. Wexler (1953) made it plausible that such an increase has been simulated by a decrease of the stratospheric transmission factor on a world-wide scale following

recent volcanic eruptions (1783, Japan, Iceland; 1815, Tambora; 1883, Krakatau; and 1912, Mount Katmai). We understand this influence much better after recent investigation on the high lifetime of radioactive particles at the middle and upper stratosphere.

(2) The well-known variations of the solar activity—in the far ultraviolet, in Roentgen rays, and in corpuscular rays—have quantitatively only a very small influence on the total energy output of the sun. In fact we do not possess any realistic and self-consistent physical model connecting the observed large responses in the upper atmosphere to solar eruptions with the tropospheric patterns and, since we ought to consider all terms of the energy balance, we are by no means justified in deriving the bulk of climatic variations from this hypothesis alone.

(3) The area and time distribution of the surface albedo—which varies between 2 per cent over sea and 89 per cent over the Antarctic—is not well known. However, there is little doubt that not only the changes of the polar ice cover but also the seasonal snow cover affect the albedo substantially, at least on a regional scale. In addition to this, the destruction of the natural vegetation and its conversion into farm land, settlements, or industry has a similar effect. Each flight over a semiarid or arid country reveals marked differences in the surface albedo between irrigated areas and bare sand or rocks, which may vary between 5 and approximately 30 per cent. Therefore the effective global radiation can be substantially—but involuntarily—altered by the activity of man.

(4) Even more variable seems to be the long-wave atmospheric counter-radiation G—increasing with the amount of absorbing constituents CO_2, H_2O, and O_3, and with the particle content of the lower atmosphere—and therefore the effective outgoing terrestrial radiation E–G. Cities and industrial areas occupy only about 0.1 per cent of the total area of the earth (Flohn, 1941). However, since industrial particles with a diameter of 1 to 10μ remain about 4 days in the lower troposphere, their effect extends to a distance of about 100 km. from their sources.

The average residence time of artificial CO_2 molecules in the atmosphere is much greater and amounts to about 5 years (Bolin and Eriksson, 1959). There seems little doubt about the fact that the CO_2 content of the atmosphere has increased from about 290 ppm before 1900 to about 330 ppm in the last decade, that is, an increase of about 12 per cent of the former value. From C_{14} investigations of living wood, Revelle and Sueß (1957) have concluded that the dilution of atmospheric carbon dioxide by C_{14}–free CO_2, produced by the burning of fossil fuels, can give only a partial explanation of the recent increase. In addition to all combustion processes, we must consider also the CO_2 production of the soil bacteria from the bare soil in cultivated areas, together with the rotting of plants, animals, and other organisms as a consequence of the increasing population. Altogether the annual production of CO_2 by industrial combustion can be estimated at 6.10^{15} gm., and the additional production by agricultural operations at 5.10^{15} gm. (Flohn, 1958).

During the dry season, the influence of large-scale—and mostly artificial—bush fires in the semihumid and semiarid tropical countries, the bird's eye view of which is so impressive, appears to be on an equal scale, as already discussed

by von Danckelman (1884). If we assume only 60 per cent of the continental areas of Köppens Aw-Climate and only 5 per cent of the area of the adjacent climates (Af, BS and, especially, Cw) to be burned annually, the resulting area can be estimated to nearly 15×10^6 km.^{2} ($\sim$3 per cent of the globe's surface). Assuming a vegetation of only 400 gm./m.^{2} (Knoche, 1937), we obtain a burned weight of 6.10^9 tons/year with a production of living CO_2 of nearly 5.10^{15} gm./year.

The particle production by all combustion processes can be estimated (Volz, 1957; cf. Lettau, 1939*a*) to be about 10^{-3} (10^{-4}) of the original mass. If we restrict our attention to dust particles with an appreciable optical effect (radius 0.1–1μ), we obtain 0.4 (0.04) gm./m.^{2} per year. Since the weight of a dust particle can be calculated from Volz's assumptions to about 6.10^{-14} gm., this is equivalent to 7×10^{12} (10^{11}) particles/m.^{2} or to 23 (2.3) particles/cm.^{2} sec. The magnitude of these values seems consistent with radiation measurements in that area (Lettau, 1939*b*).

In both cases the man-made increase of CO_2 and dust content of the atmosphere produces an increase of G with a decrease of E–G. From the first effect, Plass (1956) has computed an annual temperature rise of 0.011° C., nearly equivalent to the observed rate. In any case, we should not overemphasize the value of one single term in the thermal balance equations. The high temperatures of the European Middle Age (Flohn, 1957), the apparently worldwide fall during the late 16th century to the Little Ice Age (1680 to 1740), and the widespread warming during the period 1770 to 1810, well before the start of the industrial era, may warn us of a generalization of the CO_2 theory.

(5) The well-known global rise of temperatures (air, ocean, and soil) in the order of 0.01° C./year—except in the subantarctic latitudes (Prohaska, 1954)—shows that ΔU_L is positive, as well as ΔU_B, at least in the oceans of the Northern Hemisphere and the tropical zone, with a quite similar rise. Since the heat storage capacity of the oceans is much larger than that of the atmosphere, and since substantial changes in a geological time scale are observed, the last-named quantity and the related advective heat flux conducted by surface and lower ocean currents may play an enormous role in the mechanism of climatic change—as suggested by Ewing and Donn (1956)—but only together with other effects.

(6) One of the most sensitive quantities of the thermal balance is the relationship between U_L and U_V, the exchange of sensible and latent heat between surface and atmosphere. Variations of the actual evapotranspiration by man's activity—for example, desiccation of lakes and swamps, deforestation and reforestation, and conversion of natural vegetation into farm land—may influence climate substantially, at least on a local or even regional scale. In most continental areas, the annual averages of U_B, U_N, F, and ΔU are small as compared with U_L and U_V; if U_V is artificially changed and Q remains about constant, U_L will change in the opposite direction. Since the boundary of semihumid and semiarid areas—or, from a more practical point of view, the boundary of agricultural areas without irrigation—depends on the relationship between precipitation N_K and actual evapotranspiration E_K ($U_V = 1 \times E_K$, 1 = latent heat of water vapor), this boundary is also subject to man-made changes.

Schwerdtfeger (1955) has presented a hypothesis to the effect that the observed increase of precipitation in the Argentine pampas is due to the increase of evapotranspiration during the conversion from steppe to crop land. However, over the western prairie states of the United States no similar effect could be found (FIGURE 5; Flohn, 1957), since the fluctuations of precipitations

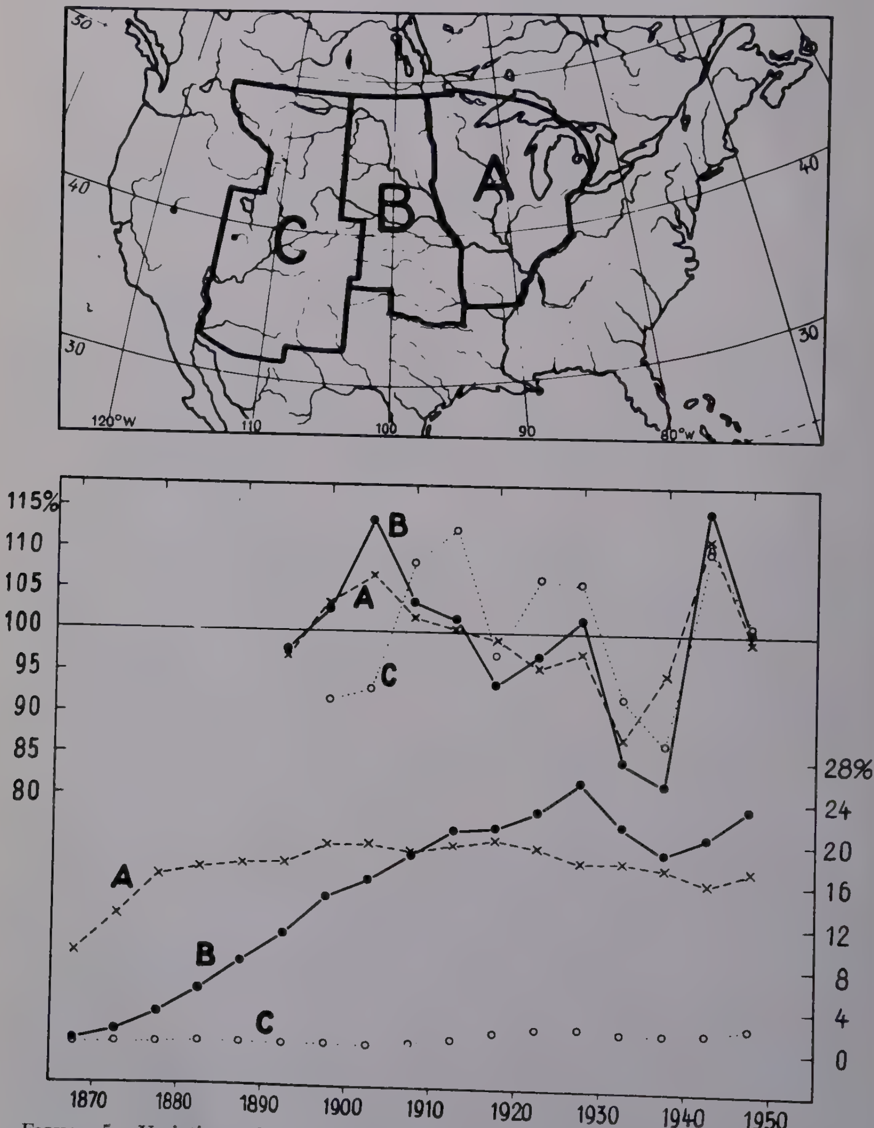


FIGURE 5. Variations of precipitation during the vegetation period (April to August, *center*, 5-year averages) and of the harvested area in per cent of the total area (*below*) of the continental United States. Key: *B* = states along the semiarid/humid boundary (North and South Dakota, Nebraska, Kansas, and Oklahoma); *A* = humid comparison area; and *C* = semiarid and arid mountain states. Reproduced by permission of the German Meteorological Service.

run more or less parallel with the adjacent territories, and the severe drought in 1932 to 1935 was followed by a substantial decrease of cultivated areas. In areas with high velocities and large vertical shear of winds such an effect can hardly be expected, due to the rather large residence time of H_2O in the atmosphere. However, at least in tropical countries, the influence of the conversion of natural vegetation into agriculture does not play a negligible role in the water balance, as recently shown by Pereira *et al.* (1961). According to these authors, the water intake of a natural bamboo forest was about 50 per cent higher than that of an adjacent cultivated catchment area.

In the tropics the water-vapor transport, which occurs chiefly in the lower troposphere, is slow and, after the destruction of the natural vegetation—especially of rain forests and swamps, but also of savannas—U_L increases with decreasing U_V and E, and the amount of precipitable water in the atmosphere must be lower than before, at least in the interior of tropical continents.

Due to the high rate of evapotranspiration and to the low velocity and vertical shear of wind, the role of local evapotranspiration (with its man-made alterations) in the water balance should be much greater over tropical continents than at higher latitudes. In the interior part of the summer rain belt over North Africa (Sudan), we can estimate the rate of turnover in the precipitation-evaporation cycle. Since in the belt with a rainy season of about 6 months the semiannual precipitation amounts to 800 to 1100 mm., on the average, to 950 mm. or 5.2 mm. per day, and the precipitable water W_L available in the atmosphere is 42 mm. (Peixoto, 1958), the average residence time of a water vapor molecule in the atmosphere in that region can be estimated at about 8 days. A similar estimate for the continental part of the equatorial rain zone in Africa (N = 1800 mm. per year, W_L = 44 mm.) yields about 9 days, as compared to the global value nearly 11 days (N = 810 mm., W_L = 24 mm.).

In this area, the role of water vapor advection from the ocean is surprisingly small, according to the values given by Peixoto (1958), since the inflow with the shallow southwest winds in the Gulf of Guinea is to a large extent compensated or even overbalanced by the outflow with the easterlies above 1 to 2 km. From these estimations, the conclusion follows that in tropical continents any large regional change of W_L by man-made alterations of evapotranspiration may similarly alter the regional precipitation in the same climatic zone as well, if not too far from the source. This is especially true for a region where the water-vapor transport in the lower troposphere is, on the average, vertically divided into two opposite directions, as in the Sudan. From this viewpoint of physical climatology, the idea of an artificial increase of evaporation in the summer rain belt of Africa, as pointed out by Bergeron (1959), should be given serious attention.

Summary

Summing up these discussions, two points should be stressed:

(1) Any theory of climatic changes based only on one factor can hardly be considered as satisfactory. A careful consideration of the atmospheric energy budget including its storage capacities may serve best as a base for discussion.

(2) Regarding water balance, CO_2 budget, and artificial air pollution, the

occurrence of man-made effects on climatic changes, even on a fairly large scale, should not be underestimated. At least in tropical continents, substantial changes in climate-producing factors are initiated involuntarily by man's activity, and we ought to visualize the danger that such effects are acting in an irreversible manner against his benefit.

References

ALDRICH, L. B. 1950. Report of the Astrophysical Observatory. Ann. Rept. Smithsonian Inst., App. **8**.

BERGERON, T. 1960. Possible man-made great-scale modifications of precipitation climate in Sudan. Geophys. Monogr. **5**: 399–401.

BERLAGE, H. P. 1957. Fluctuations of the general atmospheric circulation of more than one year, their nature and prognostic value. Verh. Med. Kon. Ned. Meteor. Inst. **69**.

BOLIN, B. & E. ERIKSSON. 1959. Changes in the carbon dioxide content of the atmosphere and sea due to fossil fuel combustion. Rossby Memorial Volume : 130.–142.

BREZOWSKY, H., H. FLOHN & P. HESS. 1951. Some remarks on the climatology of blocking action. Tellus. **3**: 191–194.

VON DANCKELMAN. 1884. Die Bewölkungsverhältnisse des südwestlichen Afrikas. Meteor. Z. **1**: 301–311.

DEFANT, A. 1921. Die Zirkulation der Atmosphäre in den gemäßigten Breiten der Erde. Geogr. Ann. **3**: 209–265.

EWING, M. & W. L. DONN. 1956. A theory of ice-ages. Science. **123**: 1061–1066; **127**: 1159–1162 (1958).

FLOHN, H. 1941. Die Tätigkeit des Menschen als Klimafaktor. Z. Erdkunde. : 13–22.

FLOHN, H. 1951. Hochgebirge und allgemeine Zirkulation. Ber. Deut. Wetterdienst US-Zone. **31**: 17–23.

FLOHN, H. 1957. Klimaschwankungen der letzten 1000 Jahre und ihre geophysikalischen Ursachen. Verhandl. 31. Dt. Geographentag.: 201–214.

FLOHN, H. 1958. Bemerkungen zum Problem der globalen Klimaschwankungen. Arch. Meteorol. Geophys. Bioklimatol. **B9**: 1–13.

HOFFMANN, G. 1960. Arch. Meteor. Geophys. Bioklim. **A11**: 474–502.

KNOCHE, W. 1937. Der Einfluß von Vegetationsbränden auf die Witterung. Meteorol. Z. **54**: 243–254.

KRAUS, E. B. 1960. Synoptic and dynamic aspects of climatic change. Quart. J. Roy. Meteorol. Soc. **86**: 1–15.

LAMB, H. H. & A. I. JOHNSON. 1959. Climatic variation and observed changes in the general circulation. Geograf. Ann. **41**: 94–134.

LETTAU, H. 1939a. Versuch einer Bilanz im Kondensationskern-Haushalt der Troposphäre im Durchschnitt für die ganze Erdoberfläche. Ann. Hydrogr. : 551–559.

LETTAU, H. 1939b. Kern- und Staubgehalt der Bodenluft und die atmosphärische Schwächung der Sonnenstrahlung über Afrika und den angrenzenden Meeren. Gerl. Beitr. Geophys. **55**: 103–127.

MANLEY, G. 1953. Climatic variation. Q. J. Roy. Meteorol. Soc. **79**: 185–207.

MITCHELL, J. M., JR. 1953. On the causes of instrumentally observed secular temperature trends. J. Meteorol. **10**: 244–261.

NAMIAS, J. 1959. Recent seasonal interactions between North Pacific waters and the overlying atmospheric circulation. J. Geophys. Research. **64**: 631–646.

NAMIAS, J. 1959. Persistence of mid-tropospheric circulations between adjacent months and seasons. Rossby Memorial Volume.: 240–248.

PEIXOTO, J. P. 1958. Hemispheric humidity conditions during the year 1950. Mass. Inst. Techn., Dept. Meteor. General Circulation Project AF 19 (604)-2242, Scient. Rept. No. 3.

PEREIRA, H. C., M. DAGG & P. H. HOSEGOOD. 1961. An intensive method of catchment-basin study. Paper presented at Int. Afric. Conf. Hydrol. Nairobi (cf. H. C. Pereira. Nature. **184**: 1768–1771 (1959).

PLASS, G. N. 1956. The carbon dioxide theory of climatic change. Tellus. **8**: 140–156.

PROHASKA, F. 1954. Bemerkungen zum säkularen Gang der Temperatur im Südpolargebiet. Arch. Meteorol. Geophys. Bioklimatol. **B5**: 327–330.

REVELLE, R. & H. E. SUESS. 1957. Carbon dioxide exchange between atmosphere and ocean and the question of an increase of atmospheric CO_2 during the past decades. Tellus. **9**: 18–27.

RODEWALD, M. 1957. Wirbelsturm-Hänfigkeit und allgemeine Zirkulation. Ann. Meteor. **8**: 167–170.

SCHERHAG, R. 1936. Die Zunahme der atmosphärischen Zirkulation in den letzten 25 Jahren. Ann. Hydrogr. **64:** 397–407. Cf. l.c. **67:** 57–67, 292–303 (1939).

SCHWERDTFEGER, W. & C. J. VASANO. 1954. La variacion secular de las precipitaciones en el Este y Centro de la Republica Argentina. Meteoros. **4:** 174–193 (cf. Meteor. Rundsch. **8:** 7–10, 1955).

VERYARD, R. G. 1956. Some thoughts on climatic change. Weather. **11:** 355–364.

VOLZ, F. 1957. Abschätzung über einige Quellen des atmosphärischen Aerosols. Geofisica pura e appl. **36:** 138–147.

WAGNER, A. 1940. Klimaänderungen und Klimaschwankungen. F. Vieweg, Braunschweig.

WEXLER, H. 1953. Radiation balance of the earth as a factor of climatic change. *In* H. Shapley: Climatic Change. **:** 73–105. Harvard Univ. Press. Cambridge, Mass.

WILLETT, H. C. 1949. Long-period Fluctuations of the General Circulation of the Atmosphere. J. Meteorol. **6:** 34–50 (cf. also Geogr. Ann. **31:** 295–315 (1949).

WILLETT, H. C. 1960. Statistical Behaviour of the General Circulation of the Northern Hemisphere, October 1945–March 1952. Sci. Rept. USWB-MIT Ext. Forecast Project.

Part IV. Glaciology, Oceanography, and Climatology

THE ICE COVER OF NORTHERN ELLESMERE ISLAND*

G. Hattersley-Smith

Geophysics Sections, Defence Research Board, Ottawa, Ont., Canada

Present Ice Cover

Northwest of the Tanquary Fiord-Lake Hazen-Alert line, rather more than one half of northern Ellesmere Island, N.W.T., Canada, is covered by ice at the present time (FIGURE 1). The relatively thin and limited ice cover of the mountains of northern Ellesmere Island is in striking contrast to the great mass of the Greenland icecap, centered many miles to the southeast. As in the case of the northernmost part of Peary Land, Greenland,[1,2] the reason for the anomaly would appear to be the very low precipitation, which averaged 14.7 cm./yr. at Alert, Canada, in latitude 82° 30′ N in the years 1951 to 1959.[3]

Recent work on the Canadian International Geophysical Year expedition showed that the interior icecap of northern Ellesmere Island reaches a height of 1800 to 2000 m., over a fairly wide area. The icecap is broken by nunataks, and flanked by mountain ranges to north and south. The highest mountain is situated near the head of Henrietta Nesmith Glacier and is 2500 m. high. The icecap rests on a very irregular rock floor, where seismic soundings showed that the ice thickness commonly varies from 400 to 800 m., in a short distance.[4] At the present time it seems that the high mountain region constitutes a reservoir of ice, accumulated in conditions more favorable to the growth of glaciers. The ice moves out, perhaps in general very slowly, toward the periphery of the high land, whence some of the ice moves in outlet glaciers down to the long fiords of the north coast, while some moves through gaps in the mountains to form piedmonts at intermediate levels, and some spills out as glacier tongues into valleys at lower elevations. Measurements of surface movement of Gilman Glacier in 1957 to 1958 gave a maximum velocity of 25 m./yr.[5] In the lower mountains of the north coastal region, intersected by the fiords, collecting grounds are too restricted and the altitudes not great enough for a general ice cover with the present scanty snowfall; glaciers occur only in locally favored areas. The plateau region between Alert and Lake Hazen is almost completely ice-free. A much greater ice cover than exists today in northern Ellesmere Island may have required a higher snowfall. Conversely a higher snowfall, such as might result from more open conditions in the Arctic Ocean, could lead to an advance of the main glaciers of northern Ellesmere Island.[6] The area is of critical interest in connection with the Ewing and Donn theory of ice ages.[7]

Former Ice Cover

It was formerly believed that the ice cover in northwestern Ellesmere Island had never been more extensive than at the present time,[8] but this view was

* Published by permission of the Chairman, Defence Research Board, Ottawa, Ont., Canada.

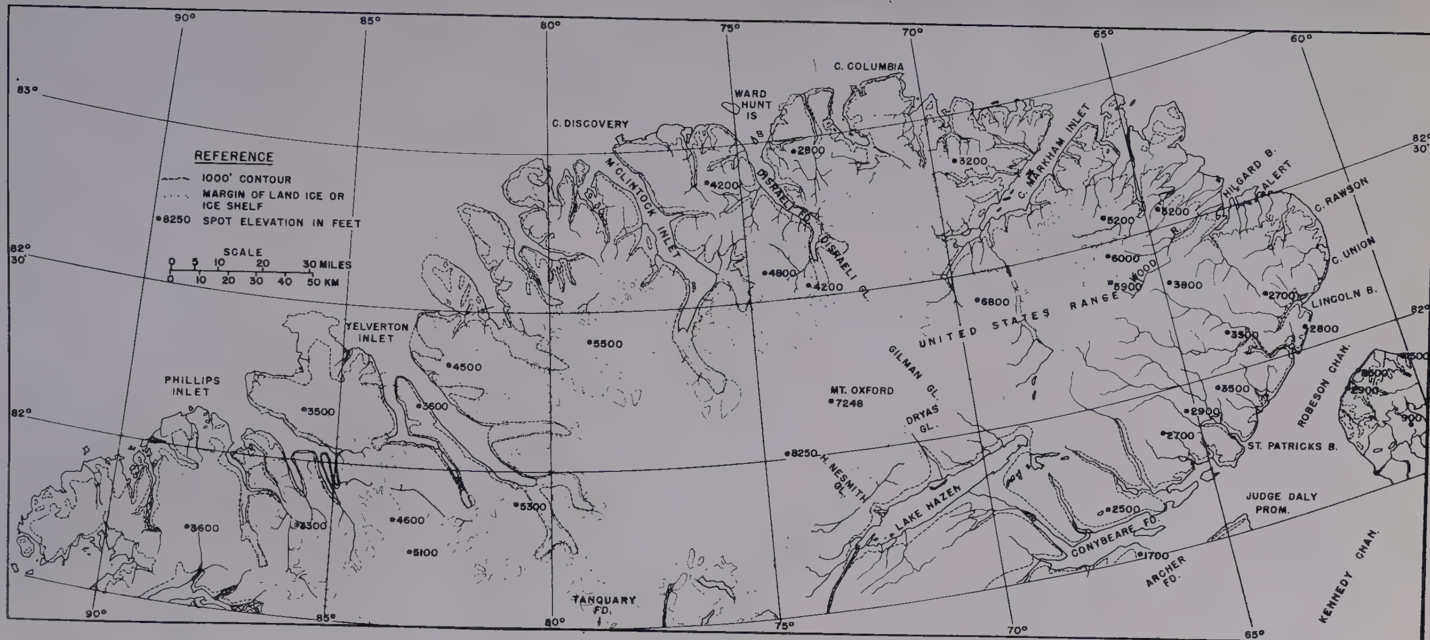


FIGURE 1. Map of Northern Ellesmere Island.

disputed by Holtedahl[9] and by Washburn.[10] Later, Troelson[11] stated that the whole of the Eureka Sound-Canyon Fiord region was covered by ice during Pleistocene time, and concluded from evidence on Stor Island in Eureka Sound that the ice cover was continuous and had its center, or centers, in eastern Ellesmere Island. Near the north coast of Ellesmere Island, Hattersley-Smith[12] found erratics from the main coast at a height of more than 300 m. on Ward Hunt Island, and glacial striae on exposed ridges at 750 m. to the south of Ward Hunt Island, where now there are only a few corrie glaciers; he concluded that the ice cover of northern Ellesmere Island had formerly been much more extensive. Taylor[13] correctly cautioned against interpreting the paucity of recorded striae and glacial deposits as evidence against widespread glaciation in a region of intense frost action and solifluction, and in a region of recent coastal submergence. He rightly pointed to the fiords, especially of western Ellesmere Island, as unequivocal evidence of former more widespread and intense glaciation. He supposed that the former ice cover of northern Ellesmere Island was almost complete, and suggested that there was a main flow of ice from Greenland. However, it seems more probable that areas in northern and southern Ellesmere Island, with their high mountains, were centers of glaciation from which glaciers moved outward to join the ice from Greenland, Labrador, and Keewatin, than that the ice invaded these areas from Greenland.[14] From the presence of erratics at a height of up to 1000 m. on the ice-free mountains between Lake Hazen and the main icecap, Smith[15] concluded that the whole of this region was covered by ice at one time, and that the flow of ice was predominantly from north to south.

The recent evidence on the extent of glaciation in northern Ellesmere Island contrasts with the evidence from the northern part of Peary Land, which is considered to have escaped major glaciation.[1]

The Ice-Free Region Between Alert and Lake Hazen

Between Alert and Lake Hazen, bounded by Robeson Channel to the east and the United States Range to the west, an extensive rolling plateau reaches an average height of 600 to 800 m.; it is devoid of ice cover at the present time except for 2 thin icecaps 15 km. north of St. Patrick's Bay. The latter lie at an altitude of about 1000 m., have a total area of less than 20 km.^{2}, and probably are receding rapidly. The surface of the plateau, which is formed by the folded and eroded Cape Rawson beds, is a peneplain of unknown age. It may be said, however, that the plateau is older than the undeformed silt and lignite deposits of Miocene or more recent age, laid down in a shallow basin on the plateau surface.[16] The plateau has been glaciated as shown by widespread erratics, and the low hills southeast of Alert give the appearance of having been molded by ice moving in a northeasterly direction.[16] According to observations made by R. L. Christie (personal communication), the distribution and height above sea level of granite erratics may indicate some encroachment near the eastern margins of the plateau by ice from Greenland. It would seem that the effect of the ice cover on the plateau has been chiefly protective rather than erosive; in fact if the plateau had suffered strong glacial erosion it is unlikely that the soft silts and lignite would have been preserved in an area where they are in part covered by a piedmont glacier at the present time.

The onset of glacial conditions caused the permanent snow line to be progressively lowered until the United States Range developed an icecap. It is assumed that pre-existing valleys in the plateau region were poorly developed and unable to channel the ice from the expanding icecap; consequently no large valley glaciers were formed; instead the ice spread out in piedmont glaciers to give a continuous ice cover. The ice overflowed into Robeson and Kennedy channels, where it probably coalesced with the Greenland icecap through ice either partly or wholly aground in the channels. Because the ice was unconfined in its movement over the plateau, it probably never reached great thickness or acquired great erosive power; its effect was to enhance the general flatness of the plateau by smoothing minor irregularities and to remove periglacial material formed by the normal processes of weathering in preglacial and interglacial times. Much of this material has been redistributed as till in depressions and valleys of the plateau region, or deposited at lower levels on the land or in the sea as outwash, following the retreat of the ice. Well-defined moraines are rare; where they have been observed, as above the west shore of Robeson Channel,[17] and at a height of about 275 m. near Alert, they indicate the final phase of small-scale alpine and corrie glaciation. Glacial striae are seldom seen, a fact that is attributed to intense frost weathering over a long period since the retreat of the ice.

The effects of the glaciation are well displayed in the drainage, which shows that the former ice cover was a westward extension of the present icecap and was not the result of invasion by ice from Greenland. Four main lines of development are seen in the drainage pattern, as follows:

(1) The major features—Archer Fiord, Conybeare Bay, Clements Markham Inlet, Robeson and Kennedy channels—follow roughly the line of strike of the rocks. These features have been modified and deepened by glacial ice from an ancient pattern.

(2) Normal strike drainage, established in the Cape Rawson beds in preglacial time, is well displayed by the streams from Wood River northward to Clements Markham Inlet, and by the main streams on Judge Daly Promontory. Between Hilgard Bay and Cape Union, rivers follow to some extent the line of strike, along which there may have been some preglacial dissection. A remnant of strike drainage also may be seen in the stream flowing into Lincoln Bay.

(3) The high land of the United States Range, above the peneplain at 600 to 800 m., is drained by consequent streams on the southeast and northwest sides. The streams, which flow across the strike of the rocks, originated as drainage channels carrying the melt water from the margins of the icecap; they lengthened their courses as the icecap retreated.

(4) The removal of ice from the fiords and channels initiated drainage from their steep truncated walls by cross-strike streams that have cut back with varying success. This type of drainage is particularly well exemplified along the western shore of Robeson Channel and on the east side of Judge Daly Promontory.

In the most recent phase, the lower courses of many of the rivers have been confused by coastal emergence of about 90 m., and streams have been diverted by movements of glaciers. On the southeast side of the United States Range, where strike drainage is least apparent, cross-strike streams flowing southeast-

ward from the plateau region presumably have swept out much of the silt and lignite deposits (where present), and have incised their courses deeply. Under present climatic conditions, in spite of the low precipitation, the short-lived but intense seasonal runoff of melt water favors the development of narrow boxlike valleys. Semipermanent masses of snowdrift ice occur at various places in these valleys. The depth of erosion of these valleys suggests that the major withdrawal of the ice occurred many centuries ago.

The Fiords of the North Coast

The fiords of the north coast provide evidence of the passage of great thicknesses of glacier ice. While the ice cover was protective chiefly in the plateau region south of Alert, farther west much higher relief resulted in higher snowfall, and greater preglacial dissection caused the ice to be channeled in its outflow from the icecap. Great outlet glaciers were formed, and the fiords were excavated out of the preexisting river valleys, running northeastward, northward, and northwestward from the height of land. Clements Markham Inlet, Disraeli Fiord, and Yelverton Inlet are typical of these fiords.

Air photographs and ground observations from a distance show that the ice-free mountains southwest of the head of Clements Markham Inlet were at one time ice-covered to a height of at least 500 m. above present sea level. The faceted spurs of the mountain ridges and the rounded knoblike hills below provide evidence of powerful glacial erosion; they indicate the movement of a glacier of considerable thickness. With the recession of glaciers from the head of the inlet, extensive delta deposits were laid down by melt-water rivers bearing large quantities of glacial outwash material. Subsequently the delta deposits, comprising layered silts and sands, have been dissected by a very intricate pattern of closely-spaced, steep-sided gullies. The deposits contain occasional thin peaty layers, with remains of saxifrage and willow, and common Arctic shell forms known from the Pleistocene and from shores today.[18] The thickness of the deposits, rising to a height of 30 m. or more above sea level, is evidence of a long period of activity by rivers since the recession of the ice.

The Disraeli Glacier appears at one time to have extended to the west of Ward Hunt Island where it excavated a submarine canyon to a depth of 874 m.,[19] extending towards the edge of the ice shelf north of Cape Discovery. The presence of the submarine canyon indicates a former thickness of at least 1000 m. for the Disraeli Glacier. Advances of glaciers down main valleys into the fiords were evidently accompanied by advances down subsidiary north-south valleys, both from the main icecap and from local icecaps developing near the coast. Piedmont glaciers were formed in the east-west valleys that follow the strike of the rocks, and coalesced to flow east and west into the fiords. The glacially-lowered cols and system of through-valleys that are evident in the landscape today resulted from these developments. Minor disorganization of drainage also resulted, but in general the glaciers appear to have followed the present river valleys, as shown by striations and alignment of glacial material. It would appear that the withdrawal of glaciers far up the fiords predated raised beach deposits that lie undisturbed at a height of 60 m. above the shore of M'Clintock Inlet, 20 km. from the entrance; shells from these deposits were assigned a carbon-14 age of about 7200 yr. by J. Lawrence Kulp.[20]

From the very striking truncated spurs above the shores of Yelverton Inlet, it is evident that the fiord was once filled with glacier ice to a height of at least 300 m. above present sea level. On the east side of the upper part of the inlet, the scarp produced by glacial scour is exceptionally well displayed above the raised beach or possible kame terrace. On the west side, glacial erosion is responsible for the high and nearly vertical cliffs and for the small hanging valleys, where streams have started cutting back since the retreat of the fiord glacier. Small glaciers in the mountains between the two arms of the fiord occupy deep-cut, V-shaped valleys, which suggest some very recent glacial recrudescence.

It is a reasonable assumption that the improvement of climate that culminated in the Climatic Optimum, 4000 to 6000 yr. ago, led to the retreat of the glaciers from the coastal region. Raised beaches were formed up to an elevation of about 100 m. during the subsequent period of land emergence; for the most part they appear to postdate the recession of the glaciers to the heads of the fiords. The ice shelves that fringe parts of the coast of northern Ellesmere Island are believed to have developed as a result of a deterioration in climate in the last 6000 yr.,[12,21] while the inland icecap is a survival of the much greater, probably complete, ice cover of earlier times. The advance of glaciers onto raised beaches and into V-shaped valleys, and the growth of ice domes over low parts of the outer coast appear, like the development of the ice shelves, to be relatively recent events that can be regarded as part of a limited reglaciation that followed the Climatic Optimum.

Glacier Regime in the Lake Hazen Area

At the present time the accumulation between an elevation of about 1450 m. and the highest part of the icecap at 2000 m. is by firn formation; it has averaged about 13 g. cm.$^{-2}$ over the last 20 yr. Between an elevation of about 1280 m. and 1450 m., interfingering of firn and superimposed ice takes place. Accumulation exclusively by formation of superimposed ice probably only occurs in a fairly narrow belt between the equilibrium line at about 1200 m. and an elevation of about 1280 m.[22] The value of accumulation/ablation at the equilibrium line is approximately 7.5 g. cm.$^{-2}$. Measurements of accumulation and ablation indicated a deficit for Gilman Glacier of about 33 million m.^{3} of water in the 1957–1958 budget year, which was about 60 per cent of the total accumulation for that year, and a larger deficit in 1956–1957; in 1958–1959 the glacier appeared to have been more nearly in a state of balance.[23] It seems clear that the regime of the glacier at the present time is negative. However, unless a deficit of the order noted in 1958 prevailed over many decades, its effect on the area of Gilman Glacier would not be pronounced for 2 reasons: (1) Gilman Glacier is tapping a vast reservoir of ice from the highest part of the icecap, where small climatic changes have least effect; and (2) the glacier rises very steeply and reaches a thickness of 400 m., 5 km. from its terminus.[24] Thus in an area where the mean annual temperature is approximately −18° C., the amount of very cold ice that melts in a relatively short ablation season causes a scarcely significant thinning. In fact the glacier has certainly not receded in recent decades, and may even have advanced slightly.

Nevertheless marginal features of Gilman Glacier and the present status of

its tributaries and of local ice masses do provide evidence of a limited glacial recession. For example, a few kilometers west of Gilman Glacier, Dryas Glacier shows signs of recent retreat of about 60 m. on the northeast side. From the presence of an Eskimo tent ring 1.7 km. east of the glacier snout, it is apparent that the glacier is close to its maximum extent of the last few hundred years.[25] The recession of local glaciers is related to the present negative regime of Gilman Glacier, of which lengthening of the ablation season and higher summer temperatures seem the most likely causes. Pit studies on the icecap at 1800 m. in 1958 gave clear evidence of warmer summers since the early 1930s.[22] On the north coast of the island, surface wastage and occasional calving of the ice shelf reflect the warming trend of recent decades.

Discussion

The past and present extent of the ice cover and the present regime of the glaciers in northern Ellesmere Island have been briefly discussed.

The annual accumulation is very low; it is appreciably exceeded by ablation in the short summer season. It is not known whether the former extensive ice cover was built up under conditions of higher accumulation or of lower ablation or whether both these conditions obtained. If the Arctic Ocean were less ice-covered than today, increased cyclonic activity, greater atmospheric humidity, more precipitation and, hence, higher accumulation could be expected. However, there is not enough evidence to support the statement that open or semiopen conditions in the Arctic Ocean were a requisite for the former much more extensive ice cover. It could be that the decisive factor has always been the amount of summer ablation, on the assumption that the present low accumulation combined with colder summers over a long period would be sufficient to produce the former extensive ice cover.

Valuable information on the climatic history of the last few hundred years would come from stratigraphic and thermal studies in a deep bore hole on the highest part of the icecap. Because the present accumulation on this icecap is one third to one quarter the accumulation on the icecap in northern Greenland, a deep core in northern Ellesmere Island might cover three or four times the time span of a core to the same depth in northern Greenland. However, because of the relative thinness of the icecap, the irregularity of the bedrock floor and the steep gradients a short distance from the center of the icecap, interpretation of annual layering in the ice probably would not be possible below about 100 m.; and certainly not at depths corresponding to deposition more than a few hundred years ago. The case is different in Greenland where the United States Corps of Engineers have successfully interpreted annual stratigraphy to great depth. For evidence on climatic conditions in northern Ellesmere Island further back than the last few hundred years, it is desirable to turn to other fields than glaciology. Investigations in various fields suggest themselves, of which the following may be mentioned. Evidence from archeological work in the Lake Hazen area indicates a trend towards slightly colder and drier conditions in the last thousand years, but the available evidence goes back no further.[21] A fruitful line of inquiry might be the dating by oxygen isotope analysis of buried glacier ice that occurs near the shore of

Lake Hazen at a distance of 15 km. from the nearest modern glacier tongue. Pollen analysis and carbon-14 dating of peat deposits near Lake Hazen, and a detailed study of the delta deposits with associated peaty layers and plant remains at the head of Clements Markham Inlet and in other areas might also add valuable data to the climatic record.

Inquiries along the above or similar lines might provide evidence of the time of recession of the last extensive ice cover of the area.

References

1. Koch, L. 1928. Contributions to the glaciology of north Greenland. Medd. om Grøn., Bd. 65, No. 2, p. 181–464.
2. Fristrup, B. 1952. Climate and glaciology of Peary Land. Geog. Rev. **42:** 87–97.
3. Canadian Meteorological Service. Unpublished data.
4. Weber, J. R. & H. Sandstrom. 1960. Geophysical methods in glaciology. Part II. Seismic measurements on Operation Hazen in 1958. Defence Research Board, Ottawa. Report D Phys. R (G) Hazen 6, 9 p.
5. Arnold, K. C. Operation Hazen Survey, 1957–1958. Defence Research Board, Ottawa. Report D Phys R (G) Hazen 5, 56 p.
6. Hattersley-Smith, G. Some remarks on glaciers and climate in northern Ellesmere Island. Geogr. Ann. In press.
7. Ewing, M. & W. L. Donn. 1956. A theory of ice ages. Science. **123:** 1061–1066.
8. Schei, P. 1903. Summary of geological results. *In* The Norwegian Polar Expedition in the *Fram*, 1898–1902. Geogr. J. **22**(1): 55–65. (map).
9. Holtedahl, O. 1917. Summary of the geological results. Report of the Second Norwegian Arctic Expedition in the *Fram*, 1898–1902. **4**(36): 27 p.
10. Washburn, A. L. 1947. Reconnaissance geology of portions of Victoria Island and adjacent regions arctic Canada. Geol. Soc. Am. Mem. **22:** xi + 142 p. (maps).
11. Troelsen, J. C. 1952. Geological investigations in Ellesmere Island, 1952. Arctic. **5**(4): 192–210.
12. Hattersley-Smith, G., A. P. Crary & R. L. Christie. 1955. Northern Ellesmere Island, 1953 and 1954. Arctic. **8**(1): 3–36.
13. Taylor, A. 1956. Physical geography of the Queen Elizabeth Islands, Canada. Am. Geogr. Soc., 12 vols. **2:** 90 p.; **3:** 99 p.
14. Hattersley-Smith, G. 1958. Review of physical geography of the Queen Elizabeth Islands, by A. Taylor. Geogr. J. **124**(Pt. 1): 111–112.
15. Smith, D. I. The glaciation of northern Ellesmere Island. Folia Geographica Danica. In press.
16. Blackadar, R. G. 1954. Geological reconnaissance north coast of Ellesmere Island, Arctic Archipelago, Northwest Territories. Geol. Survey of Canada, Paper No. 53–10, 22 p. (map).
17. Prest, V. K. 1952. Notes on the geology of parts of Ellesmere and Devon Islands, Northwest Territories. Geol. Survey of Canada, Paper No. 52–32, 15 p. (map).
18. Wagner, F. J. E. Geological Survey of Canada. Unpublished report.
19. Crary, A. P. 1956. Geophysical studies along northern Ellesmere Island. Arctic. **9**(3): 155–165.
20. Christie, R. L. 1957. Geological reconnaissance of the north coast of Ellesmere Island, District of Franklin, Northwest Territories. Geol. Survey of Canada, Paper No. 56–59, 40 p. (map).
21. Crary, A. P. 1960. Arctic ice island and ice shelf studies. Part II. Arctic. **13**(1): 32–50.
22. Hattersley-Smith, G. 1960. Studies of englacial profiles in the Lake Hazen area of northern Ellesmere Island. J. Glaciol. **3**(27): 610–625.
23. Hattersley-Smith, G. 1960. Operation Hazen glaciological studies: snow cover, accumulation and ablation. Defence Research Board, Ottawa, Report D Phys R (G) Hazen 10, 15 p.
24. Sandstrom, H. 1959. Operation Hazen. Geophysical methods in glaciology. Defence Research Board, Ottawa, Report D Phys R (G), Hazen 3, 52 p.
25. Maxwell, M. S. 1960. An archaeological analysis of eastern Grant Land, Ellesmere Island, Northwest Territories. Dept. of Northern Affairs and National Resources, Canada, Bull. No. **170** (Anthropological Series No. 49), 109 p.

THE GLACIAL HISTORY OF ALASKA: ITS BEARING ON PALEOCLIMATIC THEORY*

Thor N. V. Karlstrom
United States Geological Survey, Department of Interior, Washington, D.C.

". . . the detailed course of climatic changes has been exceedingly complicated, but beyond the intricate picture we catch a glimpse of a mathematical law which perhaps will ultimately appear quite simple. . . . To fix this system is one of the most important tasks of the future. It should be realized all around the world, but this must be done without bias, and independently within each of the various areas of work. For following a stereotyped plan leads here, as always, into a blind alley, and it is not to be expected that the course of the climatic curve has been the same in detail in all parts of the world, though in its main features it shows such a striking correspondence."

Lennart von Post (1944)

Introduction

There is no generally accepted comprehensive theory of the ice ages, nor as yet a complete understanding of the complex of physical agents responsible for historically recorded climatic trends. Mechanisms that have been proposed for past climatic change include: (1) shifting of geographic poles relative to continental land masses; (2) mountain building that provides upland surfaces for the accumulation of snow and glacial ice; (3) changes in ocean currents resulting from uplift or sea-level changes; (4) changes in the composition of the atmosphere, particularly fluctuations in CO_2 content; (5) changes in the turbidity of the atmosphere (presence of solid particles such as volcanic ash or dust) and (6) variations in solar insolation related to (*a*) the geometry of planetary movements or (*b*) intrinsic solar variations such as are indexed by changing sunspot numbers. It is possible that all these factors have played some part in the past climatic history of the earth. Which factor or combination of factors constitute the dominant element controlling the climate of the ice ages remains the major problem of Pleistocene geology and of climatology.

Current climatic theories proposed by Pleistocene geologists all require two separate mechanisms to explain the ice ages: (1) a primary geographic mechanism ascribable either to mountain building or shifting of the geographic poles, which caused a shift from warmer Tertiary climates to the colder Pleistocene climates; and (2) secondary mechanisms, either geographic or solar, to explain the repeated pulsations of Pleistocene climate from glacial to interglacial conditions.

The solar-topographic theory (Flint, 1947, 1957) postulates mountain building at the end of the Tertiary as the probable cause that initiated the colder Pleistocene climate and assumes that intrinsic variations in solar radiation, as

* Approved on March 24, 1961, for publication outside the Geological Survey.

historically indexed by the short term sunspot cycles, were sufficiently larger in the past to produce the glacial-interglacial climatic oscillations. The insolation-topographic theory (Emiliani, 1955, Emiliani and Geiss, 1957) retains mountain building as the primary mechanism of climatic cooling but substitutes geometric variations in solar insolation as worked out by Milankovitch (1941) as the likely mechanism that modulated the recorded Pleistocene climatic oscillations.

Ewing and Donn's climatic theory (1956, 1958*a*, 1958*b*, Livingston, Ewing and Donn, 1959) initiates the ice ages by a shift of the geographic north pole into the confined Arctic basin at the end of the Tertiary. The glacial-interglacial cycle is then explained as the result of a geographic mechanism involving variations in the interchange of warm Atlantic waters across the North Atlantic submerged threshold into the cold Arctic Ocean. According to their geologic-meteorologic model, shifting of the geographic pole into an open Arctic Ocean provided a polar precipitation supply for continental glaciation of the surrounding land masses. Glacial advance and concomitant lowering of sea level eventually lowered sea level to a critical level where inflow of warm Atlantic water into the colder Arctic across the submerged threshold was sufficiently curtailed to permit the Arctic Ocean to freeze over. The interglacial cycle is thus initiated by this freezeover, which cuts off polar precipitation supplies and results in turn in glacial retreat and rising sea levels. At some point the rising sea level allows sufficient incursion of warm Atlantic waters across the submerged threshold to melt the Arctic sea ice and to initiate another glacial cycle which in turn is repeated.

The Milankovitch-Pettersson theory (Karlstrom, 1955, 1957*a*) provides an elaboration of the Milankovitch geometric theory by proposing that secondary climatic oscillations, less than glacial stage in rank, are harmonically related to each other and to subordinate geometric variations in solar insolation and tidal forces as expressed in part by the Pettersson (1914) tide-generating force curve. This curve is calculated by celestial mechanics from a consideration of movements of the earth and moon about the sun. This theory also requires a primary mechanism to cool climate sufficiently at the end of Tertiary time to permit the proposed astronomic mechanisms to modulate the recorded climatic shifts; whether this involved mountain building, a shift in the geographic poles, or some presently unknown mechanism is left open.

The Antarctic coincidence theory (Fairbridge, 1960, 1961) combines the primary mechanism of polar migration to present positions within the Antarctica and Arctic regions to initiate world cooling, with the Milankovitch and sunspot theories of solar radiation variations to explain the recorded paleoclimatic oscillations. It is similar to the Milankovitch-Pettersson theory in that it proposes virtually the same series of harmonically related paleoclimatic cycles, including the cycles of ca. 550–1100–1650, and 40,000 years.

In so far as these differing climatic theories are all inferred from geologic data they may be directly tested against the increasing store of evidence relating to regional aspects of past climatic zonation and to the recorded pattern of past climatic changes. Compilation and regional synthesis of field observations on surficial deposits of Alaska made largely by United States Geological Survey

personnel over the past decade (Karlstrom and others, 1959; Karlstrom, 1960*b*) have greatly increased our understanding of Pleistocene climatic environments and the glacial history of a critically located subarctic and arctic region. Mapping and radiocarbon dating of glacial deposits in the Cook Inlet region (Karlstrom, 1955, 1956, 1957*a*, 1957*b*, 1959, 1960*a*) have provided a detailed, independently dated glacial chronology. The purpose of this paper is to summarize the pertinent Alaskan data in relation to independently dated paleoclimatic evidence from other regions and to assess their bearing on current paleoclimatic theories.

Pleistocene Climatic Zonation in Alaska

The recently completed compilation of the distribution of glacial and associated surficial deposits throughout Alaska (Karlstrom, 1960*b*) provides significant information on the nature of Pleistocene climatic zonation in the state. The morainal deposits, ranging in age from early Pleistocene to late Pleistocene and recording separate, successively less extensive glaciations, form subparallel belts flanking the alpine mountain ranges (FIGURE 1). Their regional distribution indicates that the Pleistocene glaciers: (1) fed from the same high areas that essentially comprise the modern alpine divides of the Pacific coastal ranges, the interior Alaska Range and the Arctic Brooks Range; (2) were largest near the Pacific Coast and progressively smaller northward towards the Arctic coast; and (3) were more extensive on the south slopes than on the north slopes of all the alpine ranges. This regional glacial intensity pattern, repeated during each glaciation, conforms with the distribution of existing glaciers, with regional southward inclination of the modern climatic snowline, and with present climatic zonation orographically produced by predominant precipitation supplies from the Pacific Ocean. The pattern reveals no profound regional atmospheric circulation changes throughout the period of morainal record. The recorded shifts toward more glacial climate thus appear to have been produced primarily from increased precipitation rates resulting from shifting or intensification of atmospheric circulation patterns centered, as today, in the North Pacific.

The evidence of southerly precipitation supplies for the Pleistocene glaciations in Alaska is consistent with the prevailing opinion of Pleistocene geologists that the continental glaciers in the adjoining subarctic and arctic regions were also principally supplied from southerly sources (Flint, 1957). Ewing and Donn (Livingston *et al.*, 1959), however, believe that recent published and unpublished information negates this prevailing opinion and supports their contention that the continental glaciers were supplied from an open Arctic Ocean precipitation supply and thus that the continental ice must have been thickest along the Arctic Ocean margin. This belief is not reflected in the interpretations placed on new field evidence obtained from the glaciated regions of northern Canada (Prest, 1957; Craig and Fyles, 1960; Lee, 1960). The evidence from northern Europe clearly indicates that the thickest part of the Baltic ice cap was located south of the Fennoscandian mountain chain bordering the Arctic Ocean, and not north of this topographic divide as might be expected if the Ewing and Donn (1958*a* and *b*) thesis of a polar precipitation source were correct. Recent field investigations in northern Greenland (W. E. Davies and D. B.

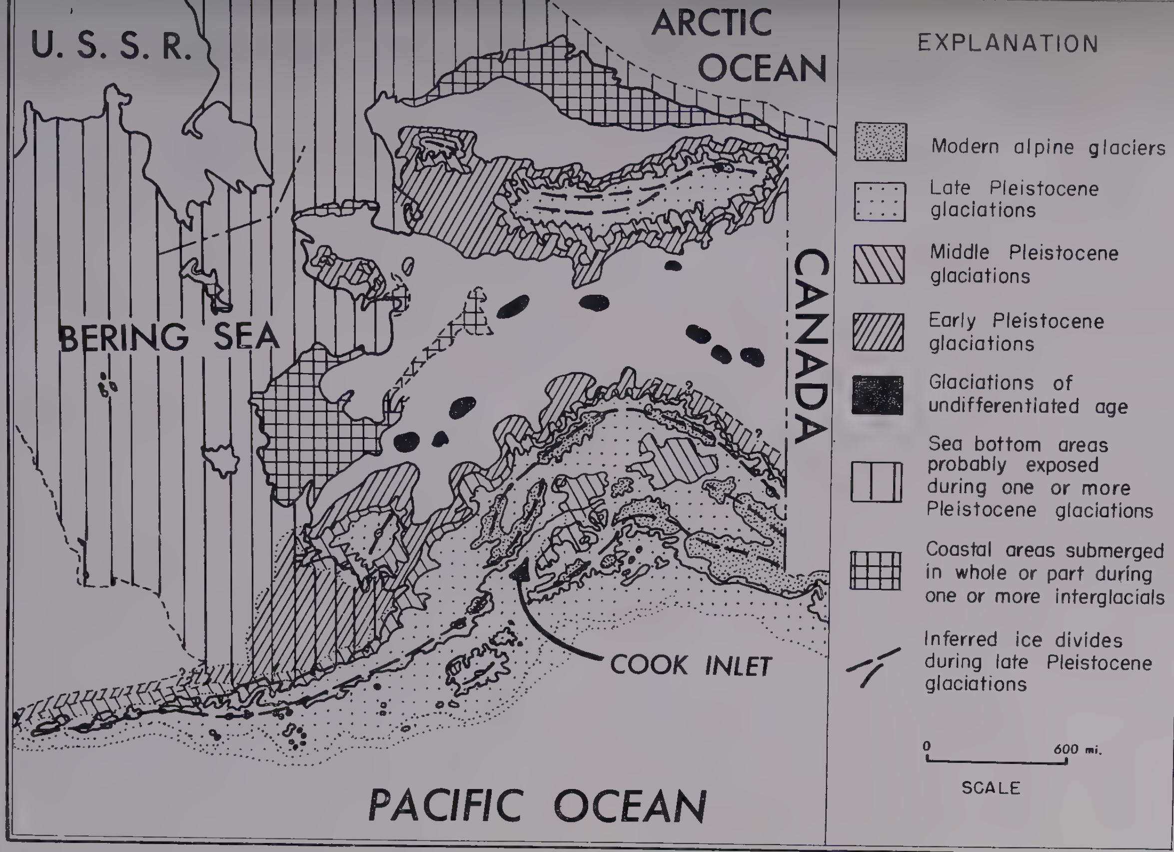


FIGURE 1. Generalized glacial map of part of Alaska reconstructed from the surficial deposits of the Alaska map in Karlstrom and others, 1959, and Karlstrom, 1960*b*. The fivefold glacial sequence of Cook Inlet is mapped as follows: the Mount Susitna and Caribou Hills glaciations as early Pleistocene; the Eklutna glaciation as middle Pleistocene; and the Knik and Naptowne glaciations as late Pleistocene.

Krinsley, oral communications, 1960) emphasize the fact that glaciations in this most northerly part of Greenland, marginal to the Arctic Ocean, were much less extensive than those to the south. It would therefore seem that the existing regional geologic data from a large sector of the Arctic Ocean coastline—Alaska to northern Europe—provide no support for the postulated polar precipitation source for continental glaciation of the Northern Hemisphere.

Quaternary Correlations

Problems of Quaternary correlation and of paleoclimatic theory are of international scope and require correlation of the available paleoclimatic evidence from all sectors of the world. The quality of published paleoclimatic chronologies varies appreciably in detail, continuity, climatic implications, and accuracy of dating. In a sound reconstruction of past climatic history it is therefore necessary to select for comparison chronologies of approximately commensurate detail that are least subject to interpretive and dating error. The chronologies presented in this paper are included on this basis, and represent the most detailed and accurately dated paleoclimatic chronologies known to the author. Because vast regions of the world, particularly in the Southern Hemisphere, are still unrepresented by detailed paleoclimatic sequences, the present sampling is disappointingly small. However, the sampling is much larger than has been available heretofore and provides much critical data bearing directly on current problems of correlation, classification, and paleoclimatic theory.

Long Quaternary Chronologies

Cook Inlet glacial chronology and North American continental correlations. The Quaternary deposits in Cook Inlet, Alaska, record five major Pleistocene glaciations and several Recent glacial advances, all separated by intervals of retreat in which alpine glaciers were probably at least as contracted as they are today (Karlstrom, 1960*a*). The late Pleistocene and Recent glacial oscillations and related depositional changes are closely dated by more than 50 radiocarbon-dated organic samples collected from critical stratigraphic boundaries. Approximate dates for older events are obtained by extrapolations controlled by roughly quantitative geologic data. Each named glaciation and glacial advance is defined in terms of moraines and associated deposits in accord with standard procedures based on Pleistocene type localities.

The major climatic oscillations, as recorded by Pleistocene and recent moraines in Cook Inlet, are graphically shown in FIGURE 2*A*. The lower boundary of the last major glaciation, the Naptowne, is dated between 46,000 and 37,000 B.C. (ca. 45,000 B.C.). This dating is based on an ionium-uranium date of 46,000 to 31,000 B.C.* for marine sediments recording a major glacio-eustatic transgression of late Knik age, and on the radiocarbon date of 37,000 B.C. (Olson and Broecker, 1959) for wood collected from stratigraphically higher deposits of early Naptowne age. Organic samples from underlying deposits of Knik and Eklutna age are all too old to be finitely dated by the radiocarbon method. The upper boundary of the Naptowne glaciation is placed ca. 3500

* Sackett, W. M., 1958, Ionium-uranium ratios in marine deposited calcium carbonates and related materials: Doctoral thesis, Washington University, St. Louis, Mo.

B.C., coincident with the culmination in late Tanya time of a marine transgression to a sea-level stand about 5 to 10 feet above present datum. As bracketed by the 2 dated higher sea-level stands, the Naptowne glaciation records a major glacial cycle of ca. 40,000 years.

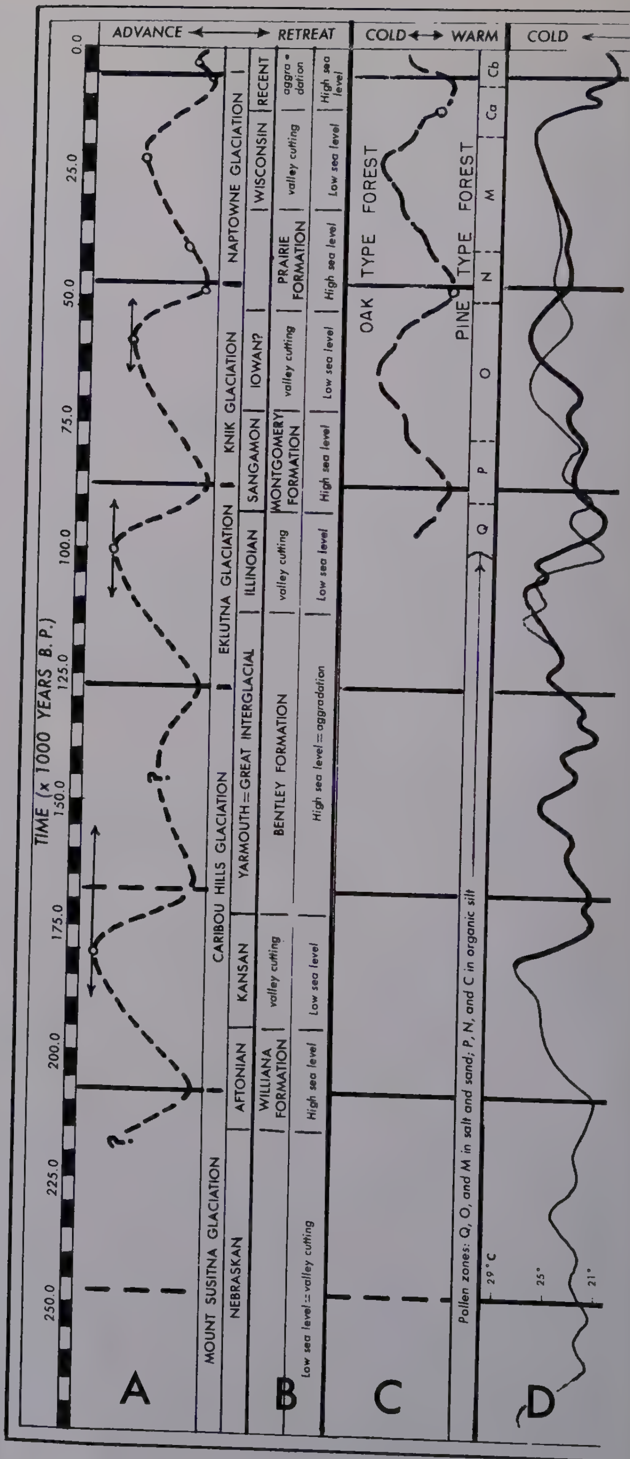
The Knik, Eklutna, and Caribou Hills glacial maxima are roughly dated respectively at 50 to 65,000, 90 to 110,000, and 155 to 190,000 years old. These dates are derived from statistical sampling of surface boulder concentrations in carefully selected sites by (1) assuming uniform rates of surface weathering, and (2) assuming that the Naptowne maximum occurred 20 to 25,000 years ago. The Mount Susitna glaciation has not yet been dated by any direct means.

The accuracy of dating falls off progressively with increasing age, but the dates applied to the pre-Naptowne glaciations are believed to be of the right order of magnitude, and to provide a first approximation to the actual age in calendar years of these glaciations in Alaska. As graphed, the culminations of the pre-Knik interglacials are placed at ca. 40,000-year intervals as extrapolations of the dated duration of the Naptowne glacial cycle. Though speculative, such a spacing is compatible with the glacial maxima ages obtained from boulder counts and conforms with other geomorphic evidence also suggesting approximately equal intervals between the Naptowne, Knik, and Eklutna glaciations, and a much longer interval between the Eklutna and Caribou Hills glaciations. By this interpretation the major Eklutna/Caribou Hills interval represents 2 interglacial culminations with an inferred intervening glaciation of subordinate intensity.

The correlation of the Cook Inlet sequence with the North American type glacial drift sequence as shown in FIGURE 2A satisfies parallel sequence and matches the long Caribou Hills/Eklutna interval with the Yarmouth interglacial, which is generally considered to mark the longest interglacial period of the midcontinental record. The Naptowne glaciation, as radiocarbon-dated, encompasses all the recognized substages of the Wisconsin (Farmdale and younger) as these are dated in midcontinental type localities (Karlstrom, 1960*a*). Thus there can be little question that the Naptowne is the temporal as well as the stratigraphic equivalent of all the Wisconsin glacial stage as this is defined from its type locality in the upper Mississippi valley.

The Knik glaciation is placed as a post-Illinoian, pre-Wisconsin event on the basis of various lines of evidence (Karlstrom, 1955, 1957*b*). A comparable event, suggested by radiocarbon dating of drift deposits in the midcontinent and provisionally referred to as the "x" glaciation by Flint and Rubin (1955) is substantially demonstrated by drift and terrace sequences worked out in Ohio by Kempton and Goldthwait (1959). Here low terraces of post-Farmdale age with soils leached through an average vertical interval of about 4 feet are separated from high terraces of Illinoian age (depth of leaching ca. 15 feet) by an intermediate set of terraces of pre-Farmdale age with leached soils averaging ca. 8-foot depths. These relative differences in soil development are consistent with the data from Alaska, which also suggest roughly equal intervals of time between the correlative Eklutna, Knik, and Naptowne glaciations.

The fivefold Pleistocene sea-level sequence of the lower Mississippi valley is shown as correlated by Fisk and McFarlan (1955) with the type midcontinental glacial sequence (FIGURE 2B). Fisk notes that the fivefold chronology of the



TIME (x 1000 YEARS B. P.)
0.0
25.0
50.0
75.0
100.0
125.0
150.0
175.0
200.0
225.0
250.0
ADVANCE
RETREAT
A
MOUNT SUSITNA GLACIATION
CARIBOU HILLS GLACIATION
EKLUTNA GLACIATION
KNIK GLACIATION
NAPTOWNE GLACIATION
NEBRASKAN
AFTONIAN
KANSAN
YARMOUTH = GREAT INTERGLACIAL
ILLINOIAN
SANGAMON
IOWAN?
WISCONSIN
RECENT
B
WILLIANA FORMATION
valley cutting
BENTLEY FORMATION
MONTGOMERY FORMATION
PRAIRIE FORMATION
aggra-dation
Low sea level = valley cutting
High sea level
Low sea level
High sea level = aggradation
COLD
WARM
C
OAK TYPE FOREST
PINE TYPE FOREST
Q
P
O
N
M
Ca
Cb
Pollen zones: Q, O, and M in salt and sand; P, N, and C in organic silt
D
29° C
25°
21°

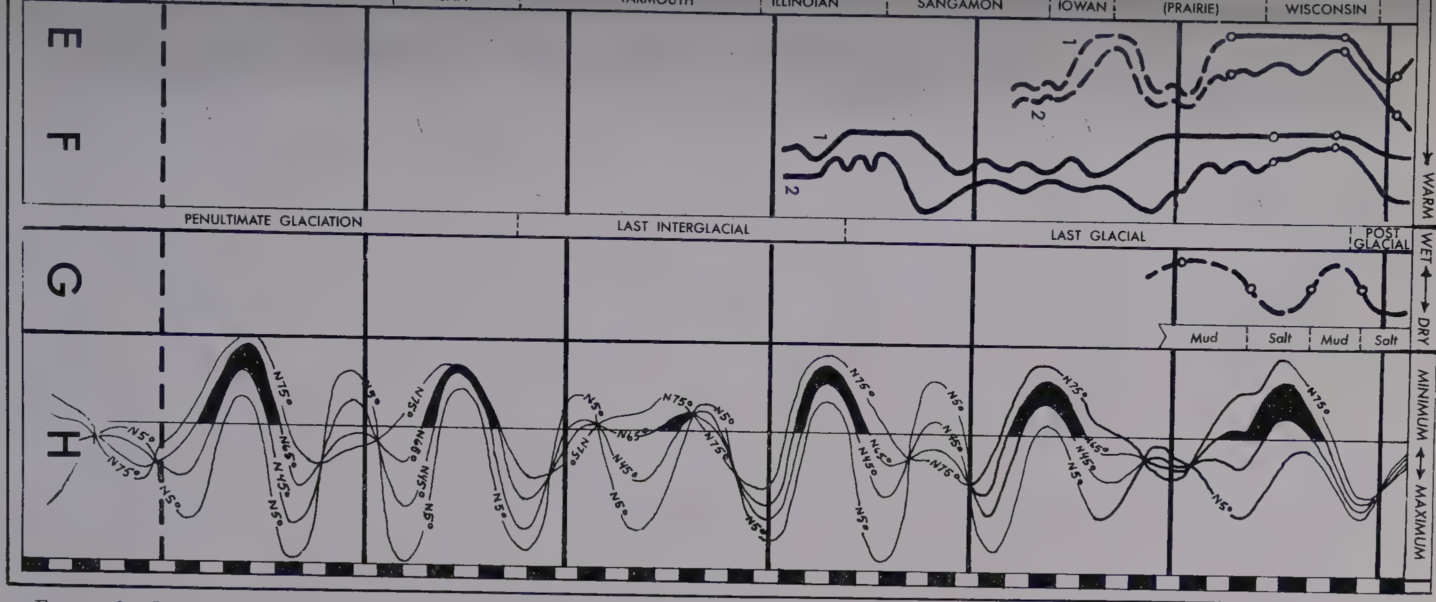


FIGURE 2. Long Quaternary chronologies:
(*A*) Generalized glacial curve of the Cook Inlet region, Alaska, and midcontinental correlations (Karlstrom, 1955, 1960*a*).
(*B*) Lower Mississippi valley chronology and midcontinental correlations (Fisk and McFarlan, 1955).
(*C*) Late Pleistocene and Recent pollen succession of coastal North Carolina (Frey, 1953, 1955).
(*D*) Generalized ocean temperature curves (Emiliani and Geiss, 1957; Emiliani, 1958). The heavy line portion of the curve represents slight revision of Emiliani's 1958 curve based on Pa-231/Th-230 dating of marine carbonate by Rosholt (Emiliani, written communication, 1961).

(*E*) Ocean temperature curves of core A-172.6 (14° 59′ N lat., 68° 51′ W long.: Ericson and Wollin, 1956; Broecker and others, 1960): Curve 1, foraminifera temperatures; Curve 2, oxygen isotope temperatures. The curves are plotted according to radiocarbon-dated sections and on the assumption of uniform deposition rates.

(*F*) Ocean temperature curves of core A180-73 (00° 10′ N lat., 23° 00′ W long.: Ericson and Wollin, 1956; Broecker and others, 1960): Curve 1, foraminifera temperatures; Curve 2, oxygen isotope temperatures. The curves are plotted according to Ericson and Wollin's (1956) time scale and continental correlations.

(*G*) Searles Lake pluvial curve. The curve is schematically reconstructed from the stratigraphic interpretation and radiocarbon dating given by Flint and Gale (1958).

(*H*) North latitude solar insolation curves for summer half years (Milankovitch, 1941). The black parts of the curve mark periods of minimal solar insolation for regions between N 65° and N 45° latitudes. Note the repeated reversals in increasing solar insolation gradients from equatorward (glacial climate) to poleward (interglacial climate).

lower Mississippi valley is not entirely compatible with the traditional fourfold glacial classification. He believes that since ocean-level fluctuations are the best markers of world-wide climatic variations, the Pleistocene sea-level chronology of the lower Mississippi valley is more significant than the traditional interpretations placed on the midcontinental drift sequence. The comparable Alaskan sequence and the new radiocarbon evidence from the midcontinental drift sequence would appear to strengthen this interpretation. Fisk favors correlation of the next to the last Pleistocene marine regression with the Iowan glaciation rather than with an early Wisconsin advance as generally accepted. In this he provides support for the original definition of the Iowan as a post-Illinoian, pre-Wisconsin glacial stage. Recent radiocarbon dating of the type Iowan drift in Iowa (Ruhe *et al.*, 1958) suggests that the type drift is much older than the post-Farmdale, pre-Tazewell loess of Iowan age of Illinois, to which it has long been correlated and, through this correlation, named by Leighton (1958). From recent stratigraphic work in Illinois, Frye and Willman (1960) recognize this probability and, to eliminate possible nomenclature confusion, they rename the loess of Iowan age in Illinois the Morton loess. Leighton (1960) and Ray (1960) however, believe that regional loess stratigraphic relations favor retention of the Iowan as a post-Farmdale substage of the Wisconsin and provide no evidence in the upper Mississippi valley for a major glacial event of post-Illinoian, pre-Wisconsin age. The pre-Farmdale age of the numerous dated wood samples collected from the base of the Iowan drift in Iowa is explained away by assuming that these samples do not represent vegetation growing at the time of the Iowan advance, but older retransported wood.

Frey's North Carolina pollen curve. The pollen record of North Carolina (Frey, 1953) is of a repeated alternation of two major forest types, a deciduous forest containing oak, and a pine-spruce forest (FIGURE 2C). This forest succession is meticulously derived from analyses of pollen and spores in lake sediments collected from a group of Carolina bays located just inland from the Atlantic coastline. The sampled lakes are located on a low-lying coastal area underlain by a granular permeable substratum and characterized by a series of terraces that are generally attributed to glacio-eustatic changes in sea level. The sampled lakes all occur below the level of the Surrey scarp (ca. 100 feet elevation) and at an elevation more than 60 feet above present datum. Little agreement exists on the sequence or ages of the Surrey scarp and associated terrace levels. As summarized by Trowbridge (1954) the Surrey terrace is dated as Sangamon by Cook, McNeil, and Vernen and as Yarmouth in age by Flint.

Frey interprets the organic silt beds (pollen zones P, N, and C) as marking intervals of higher lake levels and interglacial climate. The intercalated inorganic silt and sand beds (pollen zones M, O, and Q) are considered to mark intervals of lower lake levels and glacial climate. From a radiocarbon dating of the pollen zone N at greater than 38,000 years B.P., Frey (1955) concludes that pollen zone M must represent all of the Wisconsin glacial stage and that the older pollen zones must therefore be pre-Wisconsin in age. He speculates on the possibility that the middle horizon of his pollen sequence (pollen zone O) is

the correlative of the post-Illinoian, pre-Farmdale glaciation suggested from the midcontinent drift sequence by Flint and Rubin (1955).

From both a geographic and geologic viewpoint the environment of the Carolina bays is comparable to coastal lowland depositional basins that in Alaska record a sequence of lake-level changes (regional ground-water changes) directly determined by glacio-eustatic fluctuations in base levels. Frey's interpretation of hydrologic and climatic changes from the pollen sequences would appear to demonstrate the same direct relationship between glacio-eustatic and hydrologic changes. As correlated in FIGURE 2C on the basis of parallel sequence and radiocarbon dating, his recorded intervals of higher lake levels (higher regional ground-water levels) during pollen zones P, N, and C coincide with the recorded interglacial intervals of higher sea level. The recorded intervals of lower lake levels (pollen zones Q, O, and M) coincide with recorded glacial intervals of marine regression during the Eklutna (Illinoian), Knik (Iowan?) and Naptowne (Wisconsin) glaciations. By this correlation pollen zone P (Sangamon) is dated ca. 85,000 B.C. and pollen zone N ca. 45,000 B.C., or consistent with the greater than 38,000 years before present (B.P.) radiocarbon dating of pollen zone N. Frey's pollen chronology therefore suggests that the Sangamon interglacial seas did not reach the 60-foot level along the Carolina coastline, and that the 100-foot Surrey terrace level is more probably Yarmouth than Sangamon in age.

Emiliani's ocean temperature curve. Emiliani's ocean temperature curve (FIGURE 2D) is generalized from numerous isotope temperature curves obtained from deep-sea cores taken from the Atlantic, Caribbean, Mediterranean, and Pacific (Emiliani, 1955, 1958; Emiliani and Geiss, 1957). The upper part of the curve is time calibrated by radiocarbon dating, the time scale applied to the lower part of the curve is based on extrapolation by sedimentation rates and by comparisons with ocean core records dated by the ionium-radium measurements of Urry (1942). The suggested time scale is considered correct within 20 per cent. Emiliani concludes that the Pleistocene as defined by the continental glacial sequence (Nebraskan-Gunz to Wisconsin-Würm) falls within the last 300,000 years. He believes that the general agreement between the ocean temperature record and the Milankovitch solar radiation curve is sufficient to indicate that geometric variations in solar insolation may have caused the major fluctuations in Pleistocene climate.

Broecker *et al.* (1958) question the accuracy of Emiliani's time scale, and conclude that the Sangamon interglacial should be dated between 150,000 and 70,000 years B.P. rather than between ca. 100,000 and 75,000 years B.P. as given by Emiliani. Ericson and Wollin (1956, FIGURE 7) also suggest a comparable revision in Emiliani's chronology. They consider on the basis of somewhat dampened temperature oscillations obtained from population counts of warm- and cold-water types of Foraminifera that Emiliani's Sangamon interglacial should extend back to the beginnning of the cold oscillation that Emiliani correlates with the Kansan glaciation. Another modification in the ocean temperature time scale is given by Fairbridge (1960), who modifies the time scale slightly to conform with the variation in the rate of sediment accumulation suggested by Broecker *et al.* (1958) and notes that the resultant curve shows a

much closer coincidence with the Milankovitch solar-radiation curve. Fairbridge, however, emphasizes that only the youngest 50,000 years are proven by radiocarbon; and uncertainties remain concerning the extrapolated older dates based on measured sediment thickness and rates deduced from the radiocarbon-dated portion of the curve. All these suggested modifications are in the direction of making the Pleistocene epoch longer than calculated by Emiliani. However, Wiseman's (1958, 1959) detailed study of an equatorial deep-sea core provides no support for the Broecker *et al.* contention that ocean depositional rates necessarily differed appreciably between glacial and interglacial periods in all parts of the ocean basins.

As shown in FIGURE 2, the independently dated glacial curve of Cook Inlet and the interrelated correlations with the standard Pleistocene glacial chronology in general support Emiliani's time scale. Recent protactinium 231-thorium 230 dating of deep-sea core carbonate by Rosholt (Emiliani, written communications, Jan., 1961) confirms the general accuracy of his time scale over the past 175,000 years. The new dating requires slight revisions (shown by heavy solid line in FIGURE 2D), which contract part of the ocean temperature curve more in line with the glacial curve of Cook Inlet, rather than expanding it as suggested by Broecker and others (1958), Ericson and Wollin (1956), and Fairbridge (1960). However, the uncertainties associated with the lower end of the time scale remain so great that the apparent conformity in the early Pleistocene range between Emiliani's ocean temperature curve, the Alaskan glacial curve, and the Milankovitch solar radiation curve requires additional confirmation before it can be considered established.

The difference of interpretation placed on the ocean temperature records by Emiliani and Ericson result largely from differences in their methods of dating and of obtaining temperature changes, in their correlations to the continental glacial sequence, and in their selection of cores which, they consider, reflect with greatest fidelity past climatic changes. This may be illustrated by ocean temperature curves E and F, FIGURE 2, as reconstructed by Ericson and Wollin (1956; also *in* Broecker and others, 1960, and Ericson and others, 1961). The temperature oscillations are obtained by population studies of warm-water type and cold-water type Foraminifera in samples collected at 10 cm. intervals. The temperature curve of core 180 to 73 is shown in FIGURE 2F-1 as plotted by Ericson and Wollin (1956, FIGURE 7). The temperature curve of core A-172.6 and other cores published in Ericson and Wollin (1956) and in Broecker and others (1960) have not been plotted to a comparable time scale. These cores have therefore been time-calibrated according to the radiocarbon dated sections, assuming uniform rates of deposition. Core A-172.6 is shown in FIGURE 2E-1, and with cores 180-74 and R10-10 in FIGURE 3D. Interpolations between dated sections are indicated by solid lines; extrapolations below the oldest dated section by dashed lines. The isotope temperature curves for cores 180-73 (FIGURE 2F-2) and A-172.6 (FIGURE 2E-2) indicate good agreement between the Foraminifera and isotope-temperature methods of analysis, but with the isotope-temperature curves tending to provide a more sensitive temperature record.

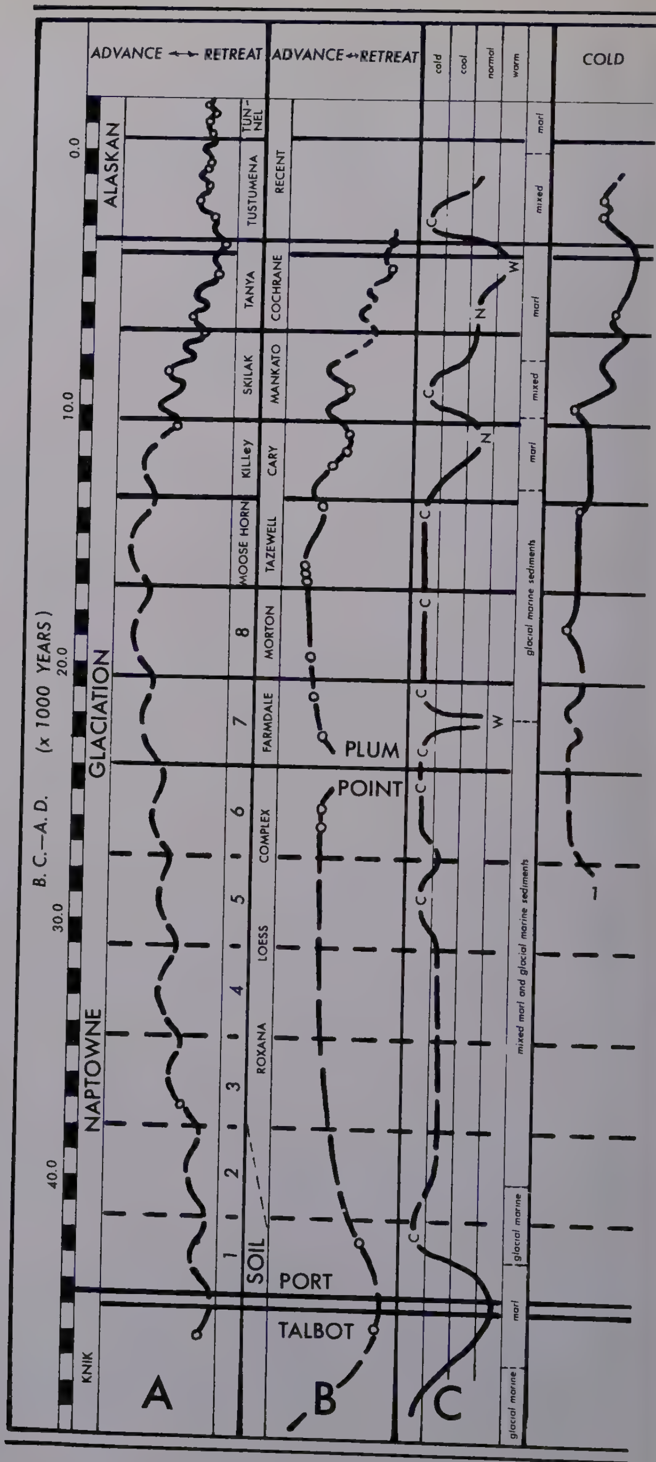
Emiliani considers that core A-172.6, which shows a marked cold period around 55,000 B.C. and a succeeding period of marked warmth centered around 45,000

B.C., is representative of the true climatic record over this time span. He believes that the evidence from this and similar cores outweighs the evidence from other cores that do not show such temperature oscillations, on the premise that in a naturally layered medium such as a deep sea deposit, any depositional disturbance or post-depositional reworking that might occur can tend only to homogenize the sediment and smooth the record (Emiliani, 1958, p. 270). Ericson, on the other hand, considers that core A-180-73 is more representative of the true climatic record and thus that the last important warming interval occurred prior to 70,000 B.C. The continental evidence is in line with Emiliani's interpretation of two periods of major ocean warming in post-Illinoian, pre-Wisconsin time. The evidence from Alaska and from the Mississippi delta region recording higher sea-level stands emphasizes the climatic importance of this event (ca. 45,000 B.C.).

Comparison of the published ocean-temperature records indicates a strong tendency for those obtained from cores in equatorial or lower-latitude sites to show temperature oscillations of lesser amplitude than those obtained from cores collected at higher latitudes. This is noted by Emiliani, who suggests that the superior detail of temperature record obtained from core 280, collected from the North Atlantic (34° 57′ N) may be due to a greater rate of sedimentation that would reduce the smoothing effect of postdepositional reworking by bottom organisms. Whatever the cause, the tendency for the higher-latitude cores to record larger amplitude temperature changes than equatorial cores may be illustrated by cores A-180-73 and A-172.6 shown in FIGURE 2 (also by cores A-180-74, A-172-6, R10-10, and P-126(5) and P-124(3) in FIGURE 3). The possibility that these latitudinal differences may in part reflect primary latitudinal differences in climatic and ocean current patterns should not be overlooked.

Searles Lake pluvial curve. The radiocarbon-dated upper part of the stratigraphic section of Searles Lake bottom sediments records two pluvial periods in late Pleistocene time (Flint and Gale, 1958). The older lake, stratigraphically represented by a mud layer (Bottom Mud) existed from before 46,000 B.P. to about 32,000 B.P. Progressive desiccation of this lake as marked by a salt layer (Lower Salt) took place between 32,000 and sometime prior to 23,000 B.P. The younger lake (Parting Mud) lasted from before 23,000 B.P. to approximately 10,000 B.P., when the last recorded period of desiccation and deposition of the Upper Salt layer commenced.

The pluvial sequence of Searles Lake as dated by Flint and Gale I have graphed in FIGURE 2G. Flint and Gale (1958, p. 712) conclude that the Searles Lake data confirm the synchronism of the last glacial with the last pluvial climate within the belt of westerly winds over the North American continent. However, if the dating of the pluvial sequence is accurate, the earlier part of the record is significantly out of phase with the other continental chronologies presented in FIGURE 2. The older lake pluvial coincides with culmination of glacial retreat, higher sea levels, and relatively high ocean-water temperatures, and the following period of desiccation (Lower Salt) coincides with a time of general glacial advance, lowering sea level and lowering ocean temperatures. It is possible, therefore, that the Searles Lake pluvial sequence may provide


B. C.—A. D. (x 1000 YEARS)
0.0
10.0
20.0
30.0
40.0
ALASKAN
GLACIATION
NAPTOWNE
KNIK
ADVANCE ↔ RETREAT
TUN-NEL
TUSTUMENA
TANYA
SKILAK
KILLey
MOOSE HORN
8
7
6
5
4
3
2
1
A
ADVANCE ↔ RETREAT
RECENT
COCHRANE
MANKATO
CARY
TAZEWELL
MORTON
FARMDALE
COMPLEX
LOESS
ROXANA
SOIL
PLUM
POINT
PORT
TALBOT
B
cold
cool
normal
warm
C
marl
mixed
marl
mixed
marl
glacial marine sediments
mixed marl and glacial marine sediments
glacial marine
marl
glacial marine
COLD

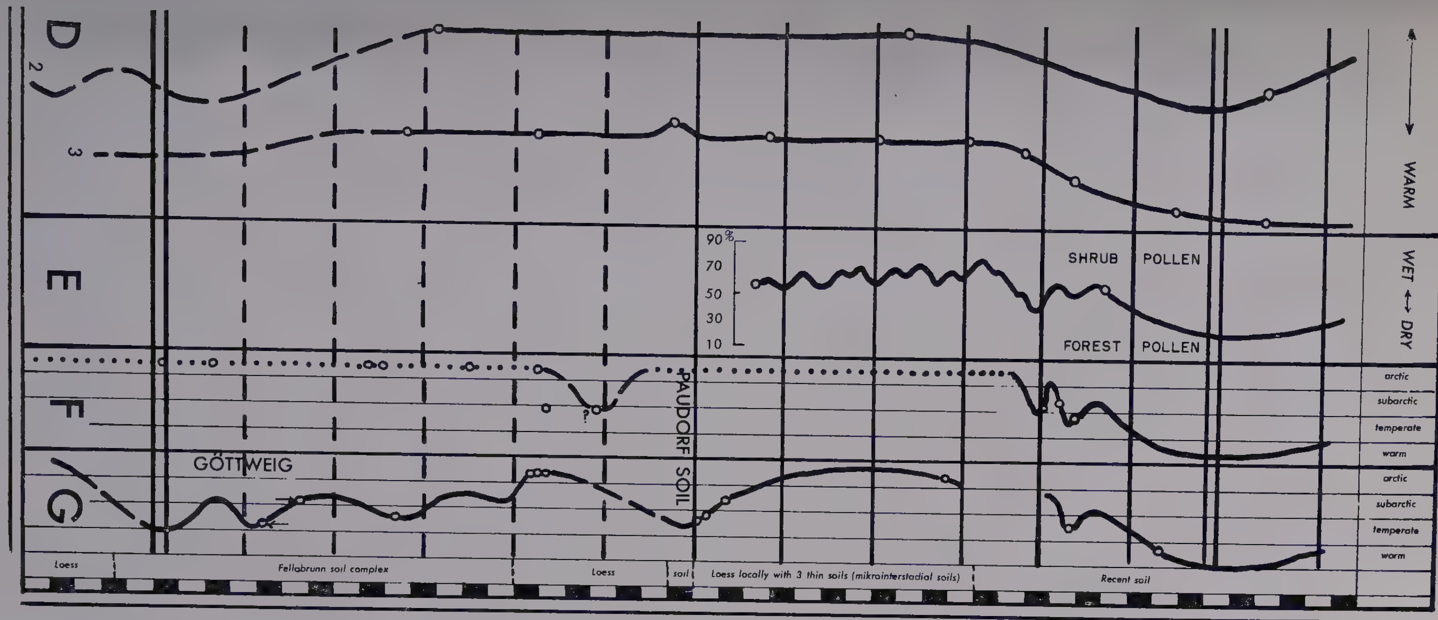


FIGURE 3. Chronologies of the last major glaciation:

(*A*) Generalized glacial curve of the Naptowne and Alaskan glaciations, Cook Inlet, Alaska, and midcontinental substage correlations (Karlstrom, 1955, 1960*a*).

(*B*) Generalized glacial curve of the Lake Huron glacial lobe (for method of reconstruction, see text and TABLE 1).

(*C*) Ocean temperature curve of North Atlantic deep sea cores P-124(3) and P-126(5) reconstructed from Hough (1953).

(*D*) Ocean temperature curves (Ericson and Wollin, 1956; Broecker and others, 1960): Curve 1, foraminifera temperatures of core R10-10 (41° 24′ N lat., 40° 06′ W long.); Curve 2, foraminifera temperatures of core A172-6 (14° 59′ N lat., 68° 51′ W long.); Curve 3, foraminifera temperatures of core 180 to 74 (00° 03′ S lat., 24° 10′ W long.). The curves are plotted according to radiocarbon dated sections and on the assumption of uniform deposition rates.

(*E*) Searles Lake pollen curve (Roosma, 1958).

(*F*) Netherlands climatic (pollen) curve (Andersen and others, 1960).

(*G*) East Central Europe climatic (loess stratigraphy) curve (Brandtner *in* Flint and Brandtner, 1961).

evidence of both interglacial and glacial pluvials as suggested by geologic evidence from other lower-latitude regions (Fairbridge, 1960). Flint (Flint and Brandtner, 1961) suggests correlation of the Lower Salt desiccation with the Paudorf and Plum Point intervals of the European and North American continental chronologies. From evidence presented below, I believe this correlation matches events of entirely different magnitudes and climatic significance.

Comparison with the Milankovitch Solar Insolation Curve

Several ambitious attempts have been made to provide a comprehensive astronomic explanation for the ice ages, but generally these have been discarded because the astronomic variations involved have been considered too small to affect terrestrial climate. The astronomic theory was most recently revived in an improved form by Milankovitch (1941), who computes changes in solar insolation as a function of the obliquity of the ecliptic, the eccentricity of the earth's orbit, and the longitude of the perihelion. Plotted in a curve, nine major minima appear in summer insolation in high latitudes in groups of two or three minima near each of four main epochs that have been correlated with the traditional four glacial stages (Günz-Nebraskan, Mindel-Kansan, Riss-Illinoian, and Würm-Wisconsin) of the Pleistocene. This astronomic theory has been vigorously supported by some European geologists (Zeuner, 1950), but has been discounted generally by American geologists (Flint, 1947, 1957). Objections raised against it have been based on geologic, meteorologic, and astronomic grounds.

The general accuracy of Milankovitch's calculations has been confirmed by von Woerkom (1953), who points out that the major configuration of the upper, middle, and northern latitude curves is dominated by the slightly more than 40,000-year obliquity cycle, and that the major configuration of the lower-latitude curves is dominated by the mean period of revolution of the line of apsides with reference to the moving equinox (precession cycle of ca. 21,000 years). von Woerkom considers that the 21,000-year cycle inferred from the Green River formation (Bradley, 1929), since it is recorded at lower latitudes, may be identified with the precession cycle, and he suggests that a climatic period of about 41,000 years might be detected in higher latitudes where the change in the obliquity must have the predominant effect on climate.

Comparison of the solar insolation curves for summer half years with the paleoclimatic record (FIGURE 2H) indicates general agreement between these theoretical considerations and the climatic record. The chronologies dominated by continental glaciations in upper latitudes (the Alaskan and midcontinental glacial sequences, the higher-latitude ocean temperature curves and the sea-level sequence) are roughly in phase with the upper north latitude solar insolation curves; whereas the Searles Lake pluvial sequence in lower-middle latitudes, as well as equatorial ocean-temperature curves, more closely parallel the lower latitude insolation curves. In all instances there is a retardation between the reconstructed glacial and pluvial maxima and the corresponding theoretical insolation minima. If, as has been proposed, such a retardation between solar insolation variations and terrestrial climate is to be expected

(Zeuner, 1950),* then the presented evidence strongly points to a causal relationship between geometric variations in solar radiation and terrestrial climate.

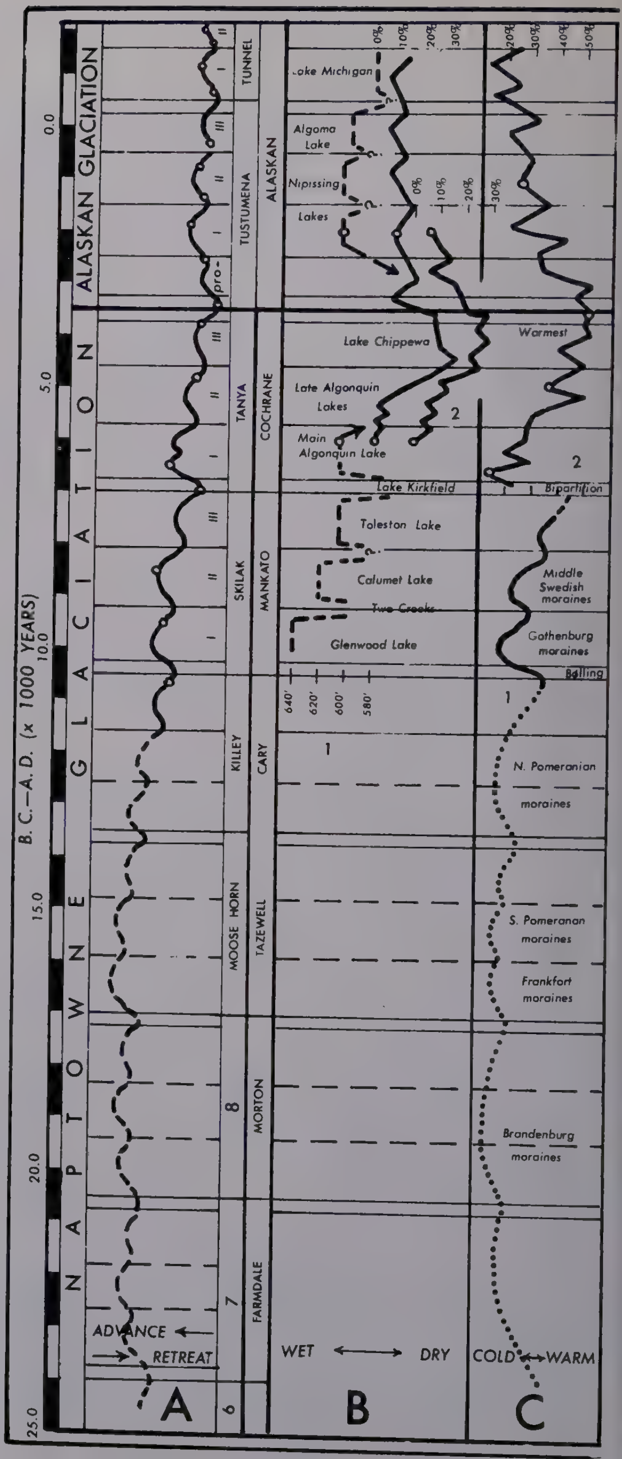
In my opinion, the ultimate geologic test of the Milankovitch theory depends on the demonstration of precessional cycle effects in middle and lower latitudes, as may be suggested by the Searles Lake data, varves in the Green River formation, and by the equatorial ocean temperature curves. In this I differ with Emiliani and Geiss (1957), who consider that the preponderance of continental ice in the upper latitudes of the Northern Hemisphere during glaciations determined secondary climatic effects on atmospheric and hydrographic circulation patterns sufficiently large to result in synchronous climatic changes across the equator and throughout the world.

One of the primary criticisms of the Milankovitch theory has been insufficiency of cause. As discussed by Willett (1953), this objection is based on the assumption that the climatic effect of geometric changes in solar insolation must relate solely to temperature changes as calculated from the effective gray body radiation by Milankovitch or from the black body radiation by Simpson. In advancing his own counter thesis of variable and intrinsic solar activity as the primary cause of terrestrial climatic changes, Willett emphasizes the irregular solar emissive activity of the highly selective sort that accompanies sunspot disturbances as the possible determinant of shifts in atmospheric circulation patterns. In view of the new paleoclimatic data suggesting correlation between major climatic phase changes and the geometric variations in solar insolation, reassessment of the solar insolation curve in terms of comparable selective mechanisms would appear to be in order. In this regard it is perhaps suggestive that when all the calculated latitudinal solar insolation curves are considered, it is seen that summer half-year intervals of solar insolation minima (equated with glacial stages) are characterized by equatorward increasing insolation gradients in the upper atmosphere; whereas solar insolation maxima (interglacial) are characterized by reversed, or poleward insolation gradients. If, in fact, these long period reversals in solar insolation gradients during summer half-years may contribute in some way to a shift in the direction of atmospheric circulation patterns from glacial (zonal pattern) to interglacial (cellular pattern), another potentially significant relationship becomes apparent. Maximum increasing equatorward solar gradients in both hemispheres are virtually contemporaneous and coincide only with the summer half-year solar minima of the Northern Hemisphere (FIGURE 6B). This in turn suggests that broad synchronism of glacial phases between hemispheres, particularly in the polar regions, might be expected to result from the total pattern of solar insolation latitudinal changes.

Chronologies of the Last Major Glaciation

Cook Inlet glacial curve and type locality correlations. The curve showing major subdivisions of the Naptowne glaciation and correlation with type locality substage events of the Wisconsin stage (Karlstrom, 1960*a*) is reproduced

* As noted by Emiliani (1955), although the new radiocarbon-based geoclimatic data support correlation with the solar insolation curve, it requires revisions in the glacial-solar minima correlations proposed by Zeuner.



ALASKAN GLACIATION
NAPTOWNE GLACIATION
B.C.—A.D. (x 1000 YEARS)
0.0
5.0
10.0
15.0
20.0
25.0
TUNNEL
TUSTUMENA
TANYA
SKILAK
KILLEY
MOOSE HORN
ALASKAN
COCHRANE
MANKATO
CARY
TAZEWELL
MORTON
FARMDALE
pro-
8
7
6
ADVANCE
RETREAT
A
Lake Michigan
Algoma Lake
Nipissing Lakes
Lake Chippewa
Late Algonquin Lakes
Main Algonquin Lake
Lake Kirkfield
Toleston Lake
Calumet Lake
Two Creeks
Glenwood Lake
640'
620'
600'
580'
0%
10%
20%
30%
40%
50%
1
2
WET
DRY
B
Warmest
Bipartition
Middle Swedish moraines
Gothenburg moraines
Bølling
N. Pomeranian moraines
S. Pomeranian moraines
Frankfort moraines
Brandenburg moraines
COLD
WARM
C

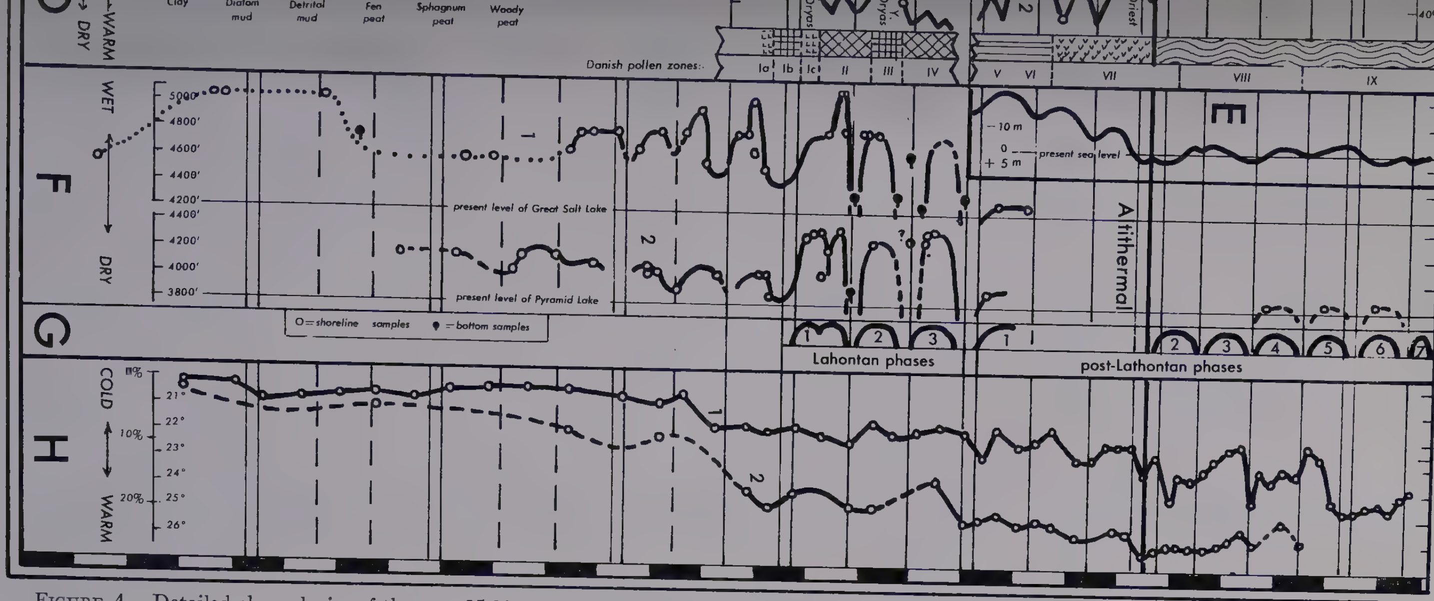


FIGURE 4. Detailed chronologies of the past 25,000 years:

(*A*) Generalized glacial curve of part of Naptowne and the Alaskan glaciations, Cook Inlet, Alaska, and midcontinental substage and stadial correlations (Karlstrom, 1955, 1960*a*).

(*B*) Climatic curves from Lake Michigan basin (Zumberge and Potzger, 1956): Curve 1, lake level sequence; Curve 2, pine pollen fluctuations from the South Haven buried bog and the Hartford surface bog.

(*C*) Climatic curves from northern Europe: Curve 1, glacial and temperature curve: the solid line part is varve-dated (E. H. DeGeer, 1957); the dotted part represents a hypothetical temperature reconstruction (Gross, 1958); Curve 2, postglacial birch pollen from a Lapland bog in northern Sweden (Lundqvist, 1956).

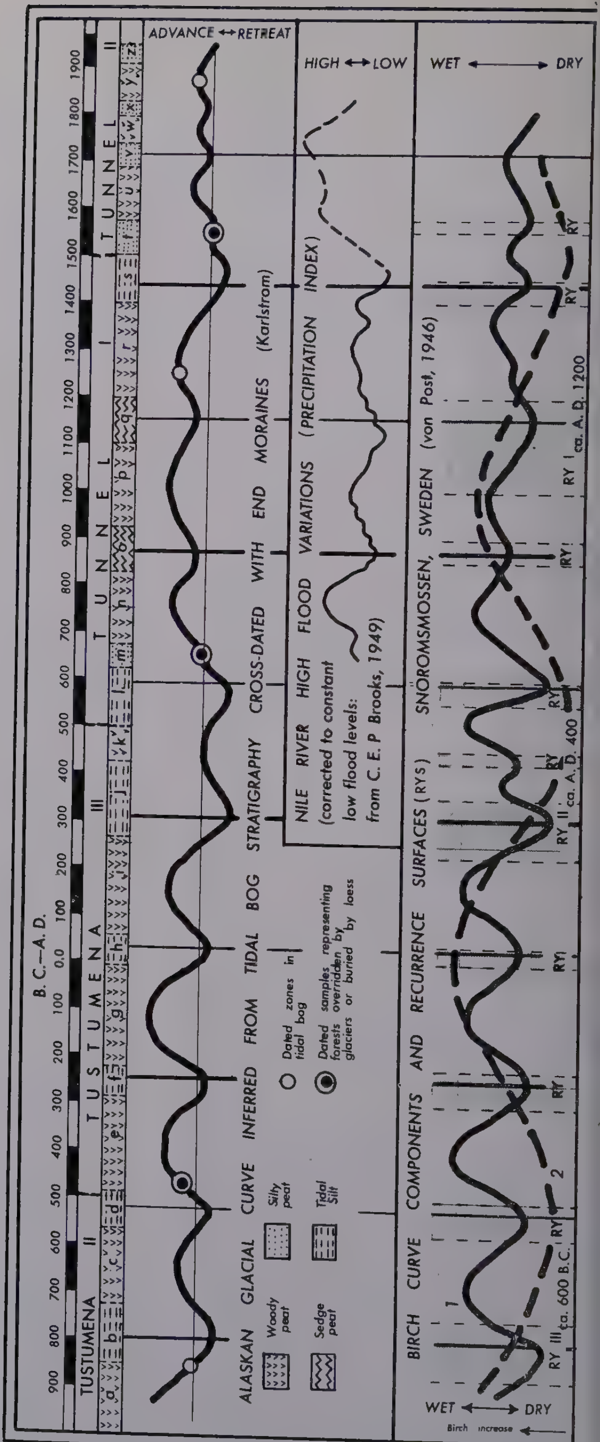
(*D*) Standard pollen chronology of northern Europe: Curve 1, birch pollen and lake stratigraphy of Bøllingsø, Denmark, representing the late-glacial pollen zones (Iversen, 1954); Curve 2, birch pollen and stratigraphy of raised bog in Emmen, Netherlands, representing the postglacial pollen zones (Zeist, 1958).

(*E*) Glacioeustatic sea level curve (Fairbridge, 1958, 1960).

(*F*) Pluvial curves of the Bonneville and Lahontan basins (for method of reconstruction see text and TABLE 2): Curve 1, Bonneville basin; Curve 2, Lahontan basin.

(*G*) The Lahontan pluvial sequence according to Morrison (1952).

(*H*) Secondary temperature oscillations in an equatorial deep sea core (1° 10′ N lat., 19° 50′ W long.: Wiseman, 1958, 1959): Curve 1, carbonate temperatures (per cent carbonate fraction greater than 500 μ); Curve 2, oxygen isotope temperatures.


B. C.—A. D.
900 800 700 600 500 400 300 200 100 0.0 100 200 300 400 500 600 700 800 900 1000 1100 1200 1300 1400 1500 1600 1700 1800 1900
TUSTUMENA II TUSTUMENA III TUNNEL I TUNNEL II
ADVANCE ↔ RETREAT
ALASKAN GLACIAL CURVE INFERRED FROM TIDAL BOG STRATIGRAPHY CROSS-DATED WITH END MORAINES (Karlstrom)
Woody peat
Silty peat
Sedge peat
Tidal Silt
Dated zones in tidal bog
Dated samples representing forests overridden by glaciers or buried by loess
HIGH ↔ LOW
NILE RIVER HIGH FLOOD VARIATIONS (PRECIPITATION INDEX)
(corrected to constant low flood levels: from C. E. P Brooks, 1949)
WET ↔ DRY
BIRCH CURVE COMPONENTS AND RECURRENCE SURFACES (RYS) SNÖROMSMOSSEN, SWEDEN (von Post, 1946)
RY III ca. 600 B.C.
RY II ca. A.D. 400
RY I ca. A.D. 1200
Birch increase

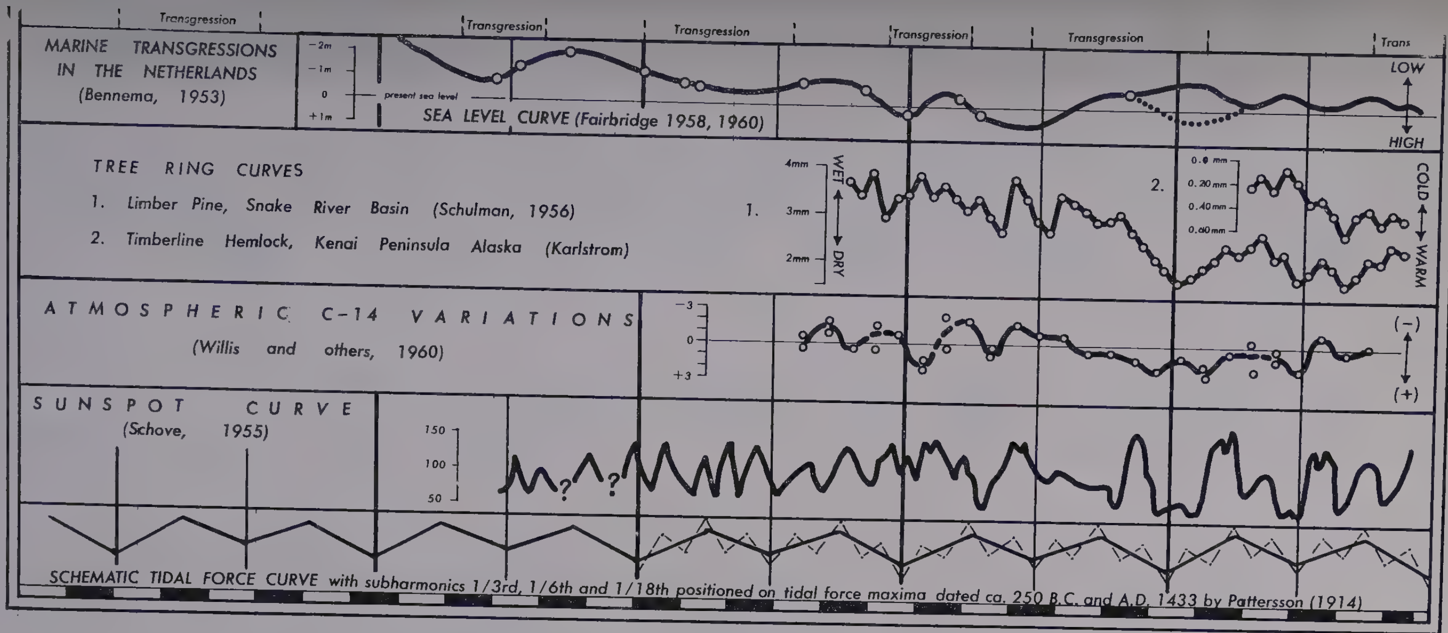
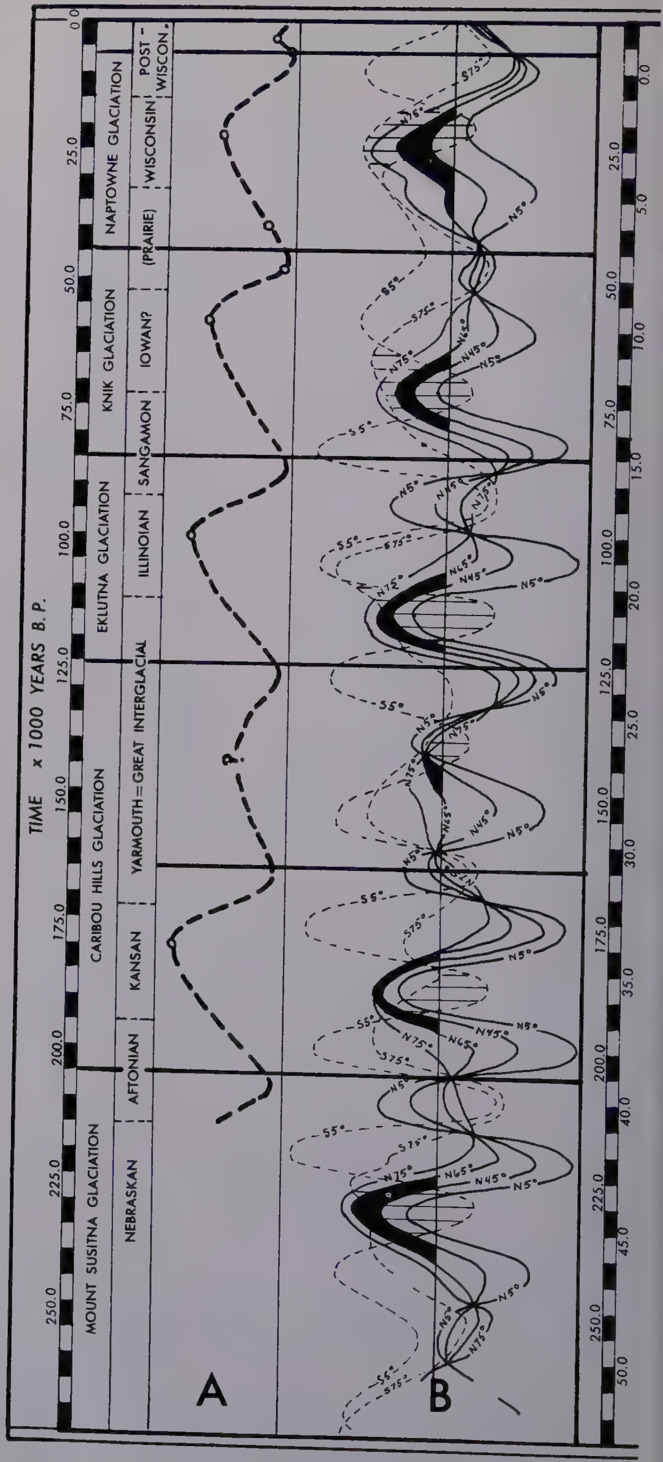


FIGURE 5. Most detailed chronologies covering the past 3000 years.



TIME x 1000 YEARS B.P.
0.0
25.0
50.0
75.0
100.0
125.0
150.0
175.0
200.0
225.0
250.0
MOUNT SUSITNA GLACIATION
CARIBOU HILLS GLACIATION
EKLUTNA GLACIATION
KNIK GLACIATION
NAPTOWNE GLACIATION
NEBRASKAN
AFTONIAN
KANSAN
YARMOUTH = GREAT INTERGLACIAL
ILLINOIAN
SANGAMON
IOWAN?
(PRAIRIE)
WISCONSIN
POST - WISCON.
A
B
N5°
N45°
N65°
N75°
S5°
S75°
5.0
10.0
15.0
20.0
25.0
30.0
35.0
40.0
45.0
50.0

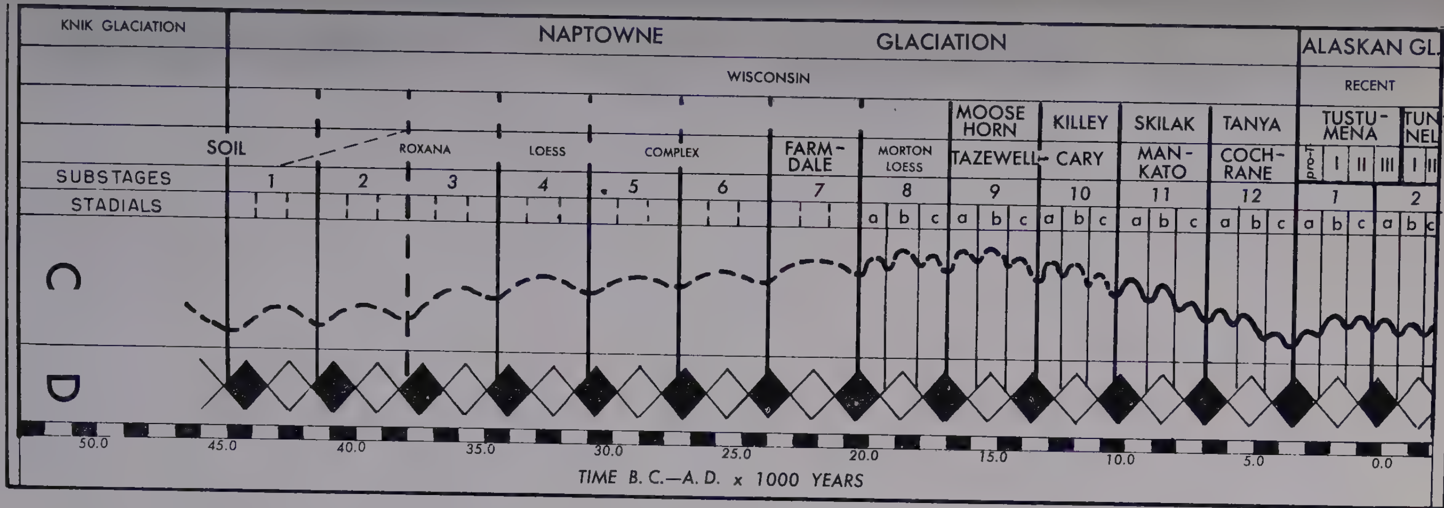


FIGURE 6. Composite paleoclimatic record compared with Milankovitch and Pettersson astroclimatic curves:

(*A*) Stage rank climatic oscillations as represented by the Cook Inlet glacial curve. The vertical lines mark interglacial culminations recorded in the Cook Inlet sequence and in one or more of the other chronologies presented in FIGURE 2.

(*B*) Latitudinal solar insolation curves for summer half years (Milankovitch, 1941). Solid lines, north latitude curves; dashed lines, south latitude curves; vertical lines mark periods, coincident with insolation minima in north latitudes, when solar insolation gradients decrease poleward in both hemispheres.

(*C*) Substage and stadial climatic oscillations as represented by the Naptowne and Alaskan part of the Cook Inlet glacial curve. Heavy vertical lines mark culminations of intraglacials; the light vertical lines, culminations of interstadials. The solid vertical lines represent boundaries recorded in 2 or more of chronologies presented in FIGURES 3 and 4; dashed vertical lines, boundaries inferred from the composite record but not yet definitively established in published chronologies.

(*D*) Pettersson tide-generating force curve (schematic) showing the major ca. 1700 year cycle as calculated from celestial mechanics (Pettersson, 1914). Alternate cycles are blacked-in to emphasize their coincidence with the dated intraglacial boundaries.

in FIGURE 3A. The solid line part of the curve is time calibrated by numerous radiocarbon-dated samples and expresses a systematic pattern of glacial oscillations with major retreats (intraglacials*) occurring every 3000–4000 years (ca. 3500 years), and important but subordinate retreats (interstadials) occurring every 1000 to 1200 years (ca. 1100 years). The dashed line part of the curve representing major oscillations of early and middle Naptowne age is reconstructed on the assumption that these oscillations were produced by the same pulsatory climatic regimen recorded for late Naptowne time. This provides the following extrapolated dates: ca. 13,500 B.C. for culmination of Moosehorn retreats, ca. 17,000 B.C. for culmination of retreat just prior to the Moosehorn maximum advance, and ca. 20,500 B.C., 24,000 B.C., and upward for earlier intraglacial retreats that presumably interrupted the general advance of glaciers to their Naptowne maximum. This reconstruction suggests 12 main subdivisions or substages of the Naptowne (Wisconsin) glaciations between the boundary dates of ca. 45,000 and 3500 B.C.

The radiocarbon-dated type locality deposits of Wisconsin substage events—Farmdale, Iowan (Morton loess of Frye and Willman, 1960), Tazewell, Cary, Mankato, and Cochrane—demonstrably fall within the respective intraglacial time boundaries derived from the Cook Inlet glacial sequence. The Tazewell-Cary boundary has yet to be definitively dated within its type locality in Illinois, and the possibility remains that either the Tazewell-Cary boundary as defined by the Minooka moraine, or the early Tazewell (Shelbyville)-late Tazewell (Bloomington) boundary as defined by the Bloomington end moraines in Illinois may be the equivalent of the Moosehorn intraglacial culmination dated ca. 13,500 B.C.

There is a general paucity of stratigraphic detail and definitive radiocarbon data in the pre-Farmdale part of the last major glacial advance. The multiplicity of events recorded and the radiocarbon ages of the Roxana loess as discussed by Frye and Willman (1960) are, however, generally consistent with its correlation with the early Naptowne sequence, and the correlation of the underlying soil developed in pre-Wisconsin drift with the Knik-Naptowne boundary at ca. 45,000 B.C. Estimating on the basis of loess thicknesses, Frye and Willman suggest age ranges for the soil of about either 50,000 or 70,000 years ago.

Leighton (1960) does not recognize the Roxana loess as a valid stratigraphic unit, and he also discounts the validity of the radiocarbon ages obtained from gastropod samples collected from the Roxana loess. Leighton redefines the Farmdale to include all events between the loess of Iowan age (Frye and Willman's Morton loess) and the Sangamon interglacial, and recognizes no major post-Illinoian, pre-Wisconsin glaciation. In my opinion the significant stratigraphic definition resulting from radiocarbon dating is that the larger peaty or woody horizons intercalated within the midwest loessal sequence and associated with the Farmdale loess date contemporaneous with intraglacial intervals as

* Little agreement exists in the literature on the terminology applied to glacial events of differing rank. Intraglacial and interstadial are used in this paper in the general sense as proposed by Leighton (1958): intraglacials = important retreats separating events of substage rank; stadials = smaller units of time represented by the major oscillations within a glacial substage.

recorded from other regions. Thus they may be interpreted as significant stratigraphic boundaries representing periods of nondeposition or slow deposition of loess (glacial retreats) separating periods of rapid loess deposition accompanying glacial advance.

Karlstrom (1956) marshals climatic evidence in support of the interpretation that the Cochrane represents a substage of the Wisconsin age. Leighton (1960) places the end of the Wisconsin at the climatic change that halted the Valders advance about 9000 B.C. This difference of interpretation results from differences in climatic reference. My subdivisions between glacial events are placed at culminations of glacial retreats, not at the beginnings of these retreats. This is done on the premise that such a convention emphasizes the most significant divisional boundaries in an oscillatory climatic sequence, provides the most convenient method of systematic subdivision of glacial periods into their subordinate components, and cuts the nomenclature problem in half by eliminating the necessity of naming both the advancing and retreating phases of each recorded glacial fluctuation. The practicality of this approach can perhaps best be judged from the nature and complexity of the climatic evidence discussed in this paper.

Chronology of the Lake Huron glacial lobe. A relatively closely radiocarbon dated chronology of the advance and retreatal phases of the Lake Huron glacial lobe of Wisconsin age is now available. Results of the pertinent radiocarbon analyses are discussed by Flint and Rubin (1955), Flint (1956), Goldthwait (1958), Dreimanis (1960), and Terasmae and Hughes (1960). Flint and Rubin's suggestion that the radiocarbon age of numerous wood samples collected from the base of the last drift sheet from Canada to Ohio record a progressive glacial advance beginning before 30,000 B.C. and terminating about 18,000 B.C. is considered by Goldthwait to be substantially demonstrated by the more recent radiocarbon dating in Ohio. Wood collected from outermost end moraine deposits in Ohio record maximum extension of the Miami sublobe around 19,000 to 18,000 B.C. (samples W-37, W-304), and of the Scioti sublobe at a somewhat later time, ca. 17,000 to 16,000 B.C. (samples Y-448, W-91, W-331). No major temporal gaps in the sequence of radiocarbon dates are revealed, and Goldthwait concludes that retreats during the general advance of the Wisconsin ice lobes into Ohio, if there were any, were short-lived.

Radiocarbon-dated stratigraphy on the Canadian side of Lake Erie is interpreted by Dreimanis to record three main Wisconsin advances, the middle one of which may not have reached south into Ohio. An "early" Wisconsin advance was separated from a "middle" Wisconsin advance by a retreatal interval in which the Port Talbot organic deposits were laid down. The latter half of the Port Talbot interval is dated between 45,500 B.C. and 42,250 B.C. Retreat from the subsequent "middle" Wisconsin advance, and readvance of the "main" Wisconsin ice is dated between 26,000 and 22,000 B.C. on the basis of the radiocarbon age of transported wood incorporated in basal till near Plum Point. The age of these wood samples bracket the age of the basal organic layer associated with the Farmdale loess in Illinois, and Dreimanis suggests that this wood represents an interstadial which he names Plum Point. However, in the absence of other stratigraphic evidence in Ontario for this inter-

stadial,* it is evident that the transported wood can be equally well explained by the overriding of a forest by a progressively advancing glacier without the necessity of postulating a preceding retreatal oscillation of major magnitude.

A complex series of ice-front oscillations accompanying general retreat of the Wisconsin ice sheet back into Canada from positions of maximum extension in Ohio is recorded by a multiple system of moraines and associated glacial lake shore lines in the Great Lakes region. Not all of these events have been radiocarbon-dated, nor is there complete agreement in detail concerning the intercorrelation of events from one region to another (Zumberge, 1960).

No glacial advance can be any older than the age of the youngest vegetation overridden by the glacier, nor any younger than the age of the oldest vegetation that began to grow in the area following retreat. The Lake Huron glacial lobe curve (FIGURE 3B) is reconstructed on this principle, and the critical dated samples used in its reconstruction are listed in TABLE 1.

Comparison of the Lake Huron glacial lobe curve with the Alaskan glacial curve and correlative midcontinental type locality events supports the following conclusions: (1) the Port Talbot interval correlates with the pre-Roxana buried soil of pre-Wisconsin age in Illinois and with the late Knik high sea-level stand in Alaska; (2) the maximum extension of the Miami sublobe took place during the time the Morton loess was being deposited in Illinois, whereas the maximum extension of the Scioti lobe was contemporaneous with the maximum extension of Wisconsin (Shelbyville) ice in Illinois; (3) the readvance dated ca. 13,500 B.C. in Ohio (Y-450) was contemporaneous with the maximum Killey advance in Alaska, and perhaps also with the Bloomington advance in Illinois in so far as the Havana terrace, dated 13,650 B.C. (W-381) has been directly correlated with the Bloomington moraine (Wanless and Leighton *in* Rubin and Alexander, 1958); (4) the Port Huron moraines and Skilak I advance in Alaska correlate with the Bemis-Altimont moraines of the type Mankato locality in Iowa as discussed by Karlstrom (1960*a*)† and the Valders with the Skilak II advances.

The difficulty of working out detailed climatic chronologies from geomorphic and stratigraphic studies of glacial drift units characterized by numerous hiatuses, and based on examination of scattered exposures, is generally recognized. More continuous depositional records such as may be represented by lake, ocean, and bog sediments and from other types of bedded deposits offer the advantage of greater continuity. As will be shown, independently dated rec-

* According to definitions used in this paper, the Plum Point interval would be an intraglacial.

† Correlation of the Skilak I with the Mankato maximum is based on numerous C-14 samples from Iowa that are internally consistent with stratigraphic position in dating the Bemis-Altimont moraines of the type Mankato Des Moines lobe drift between ca. 10,500 B.C. and 9700 B.C. More recently, published dates of the Britt forest in Iowa as interpreted in Rubin and Alexander (1960, p. 144) suggest that the Bemis-Altimont moraines are Cary in age, predate rather than postdate ca. 10,500 B.C., and thus that the Chicago dates (C-596, C-653) obtained from samples at the base of Bemis till must be wrong. However, in so far as the Bemis-Altimont moraines of the contiguous James lobe also radiocarbon date between ca. 10,500 B.C. (samples Y-595, Y-452) and 9500 B.C. (sample W-542) there is a strong suggestion that either the stratigraphic interpretation placed on the Britt forest, or its radiocarbon age may be in error. Deevey and others (1959) note the agreement in the Chicago and Yale samples listed above and consider that they date deposits of the Mankato substage.

TABLE 1

RADIOCARBON SAMPLES DATING THE ADVANCE AND RETREAT OF THE LAKE HURON GLACIAL LOBE DURING WISCONSIN TIME

Sample No.	Age (B.C.)	Description	Source
Gro-2597 Gro-2601	45,550 ± 250	Dates Port Talbot gyttya	Dreimanis (1960)
Gro-2580	42,250 ± 1500	Wood in "middle" Wisconsin till	Dreimanis (1960)
L-185B W-177	26,250 ± 1500 25,500 ± 1200	Larchwood in "main" Wisconsin till near Plum Point, Ont.	Dreimanis (1960)
L-217B	22,650 ± 1600	Spruce wood in "main" Wisconsin till near Plum Point, Ont.	Dreimanis, Goldthwait
W-71	22,650 ± 800	Wood underlying "main" Wisconsin till near Cleveland, Ohio	Goldthwait (1958)
Y-449	21,050 ± 850	Wood in basal "main" Wisconsin till near Columbus, Ohio	Goldthwait (1958)
W-117	19,650 ± 1000	Wood in basal Wisconsin till near Harrisburg, Ohio	Goldthwait (1958)
W-448	16,550 ± 420	Wood in till of Cuba end moraine near Cuba, Ohio	Goldthwait (1958)
W-91	16,100 ± 400	Wood in basal "main" Wisconsin till near Chillicothe	Goldthwait (1958)
W-331	16,020 ± 400	Wood in till at Wisconsin drift border near Anderson, Ohio	Goldthwait (1958)
Y-450	13,610* ± 230	Wood in surface till near Darrtown, Ohio: dates a glacial readvance	Goldthwait (1958)
W-198	12,350 ± 450	Wood from deposits overlying Wabash moraine of late Cary age near Edon, Ohio: postdates retreat from Wabash moraine	Goldthwait (1958)
W-33	11,650 ± 600	Organic samples collected near Pakertown, Ohio: dates low Lake Arkona of late Cary age	Goldthwait (1958)
W-240	10,850 ± 250	Wood collected near Bellvue, Ohio: dates a phase, probably an early phase of glacial Lake Whittlesey which formed during the Port Huron advance	Goldthwait (1958)
Two Creeks forest samples	9,350 ± 130 (weighted av. of 8 samples)	Collected from Two Creeks forest in Wisconsin: dates retreatal phase between Port Huron and Valders advances	Broecker and others (1960)
Y-222 W-136 W-176	4,780 ± 200 4,430 ± 350 3,350 ± 300	Basal peat collected in Cochrane area, Ont.: postdate Cochrane readvances and final retreat of ice sheet north into Hudson Bay region	Karlstrom (1956)

* Sample Y-450 shown as listed in Deevey and others (1959); Goldthwait (1958) lists its age incorrectly as 16,560 ± 230 years B.P. (14,610 B.C. ± 230).

ords of this type are in general agreement with the chronological interpretations placed on the continental record of the last major glaciation.

Ocean temperature curves. Curve C in FIGURE 3 represents ocean temperature changes inferred from the North Atlantic cores P-126(5) (48° 38′ N lat., 36° 01′ W long.) and P-124(3) (46° 03′ N lat., 43° 23′ W long.) as discussed by Hough (1953). Temperature changes are inferred from the character of the sediments and foraminiferal types. Foraminiferal marls are assumed to represent generally warmer periods; the sediments intermediate in texture and composition to the marls and glacial marine sediments, periods of intermediate temperatures. The fossil temperature indicators are shown in proper stratigraphic position by means of the letters C (cold), W (warm), and N (normal). The core was dated by Urry using "the percentage of equilibrium method" for uranium, ionium, and radium (Piggot and Urry, 1942). Hough correlates this curve with south Pacific core evidence, also dated by the Urry method, and considers that the agreement is sufficient to give considerable confidence in the dating method at least in its upper range. Ericson and Wollin (1956) also conclude that another core, P-136, as dated by the Urry method, is consistent with their radiocarbon-dated cores.

The main significance of the Urry-dated North Atlantic cores P-126(5) and P-124(3) is that they clearly record major periods of warming contemporaneous with the beginning and end of the last major glaciation ca. 45,000 B.C. and ca. 3500 B.C. In addition, the record is sufficiently sensitive to record short intervals of ocean warming coincident with some of the dated intraglacial intervals of the continental glacial record. An apparent discrepancy in the core record lies in the dating of a warm interval ca. 22,000 B.C. or apparently contemporaneous with the Farmdale loess (cold) as dated in the continental sequence and as represented by cold ocean temperatures in the other deep sea cores presented in FIGURES 2 and 3.

It is perhaps significant that cores P126(5) and P-124(3)—collected from higher northern latitudes—provide a somewhat more sensitive temperature record than cores collected from lower latitudes. In keeping with this suggestion of latitudinal control, the most northerly Atlantic core (41° 24′ N lat., 40° 06′ W long.) analyzed by Ericson also provides his most detailed temperature record. As closely dated by radiocarbon, the temperature shifts in core R10-10 (curve D-1, FIGURE 3) closely coincide with intraglacial boundaries between the Farmdale, Morton, Tazewell-Cary, Mankato, Cochrane, and Alaskan substages as established from the North American glacial record.

Ericson and Wollin (1959) assess the deep-sea core evidence from the Arctic, Atlantic, Caribbean, and Gulf of Mexico, and conclude that the synchronism of temperature changes recorded from the various ocean basins suggests that the prime climatic cause was world-wide in effect and therefore probably extraterrestrial. They consider that change of conditions in the Arctic sea was probably accentuated by greater influx of warm Atlantic water as eustatic sea level rose following the melting of the continental glaciers; but they look on this increased flow into the Arctic basin as merely a consequence, not a cause, of a global climatic change brought about probably by some extraterrestrial cause. On the other hand, Broecker *et al.* (1960) consider that Ericson and Wollin's deep-sea core data demonstrate an abrupt unidirectional warming

about 9000 B.C. in support of the thesis that Pleistocene climatic oscillations were primarily determined by eustatic sea-level controlled rates of warm Atlantic water inflow into the Arctic basin.

It is quite clear from the paleoclimatic evidence presented in this paper that this postulated unique period of warming ca. 9000 B.C., which is believed to mark the last time of Arctic Ocean freeze over and consequent abrupt shift towards climatic warming, did not occur. A warming trend ca. 9000 B.C. is clearly recorded in many types of records. Just as clearly, however, this event marks but a subordinate oscillation superposed on a major trend towards warming that began generally between ca. 20,000 B.C. and 15,000 B.C. and culminated ca. 3500 B.C. One of the major weaknesses of the Ewing and Donn climatic theory thus is that it provides no mechanism to explain the multiplicity of significant climatic oscillations geologically recorded in the late Pleistocene and Recent paleoclimatic record.

Pollen and loess curves from Europe. Curves F and G in FIGURE 3 represent temperature changes as inferred respectively from pollen evidence in the Netherlands (Anderson and others, 1960) and from loess stratigraphy in eastern Central Europe (Brandtner, 1950, 1954; Flint and Brandtner, 1961). Both curves are time calibrated by radiocarbon-dated samples marked by circles in FIGURE 3. Qualitative temperature fluctuations in the Netherlands curve are inferred from disconnected pollen sequences, those from east Central Europe from buried soils intercalated in a fossiliferous loess sequence including the Krems soil formation (Eem interglacial and two younger intervals of warming), the Fellabrunn soil complex (Göttweig interval), the Paudorf soil and, locally (South Moravia), three post-Paudorf, pre-Alleröd buried mikrointerstadial soils (Klima, 1958, Pelíšek, 1958).

The climatic curve derived by Brandtner (Flint and Brandtner, 1961) from the loess sequence is similar to the North American sequence in that the Göttweig interval dates contemporaneous with the late Knik (Port Talbot) interval, the Paudorf soil with the lower peat associated with the Farmdale loess, and the Alleröd with the Two Creeks forestbed. The pre-Paudorf part of the curve in general reproduces the same pattern of intraglacials as that inferred from the North American glacial sequence, the dated soils coinciding, within limits of dating error, with four of the six inferred intraglacial boundaries. However, additional radiocarbon data and stratigraphic work are necessary to define with greater precision the number and age of these pre-Paudorf soils.

The Netherlands chronology is based on radiocarbon dating of a post-Eem pollen sequence from which two interstadials are inferred and dated 51,000 to 57,000 B.C. (Brørup) and 64,000 B.C. (Amersfoort); a series of isolated peat layers intercalated in coastal sediments whose pollen spectra are largely of non arboreal pollen and are positioned within the chronology solely on the basis of their radiocarbon age; and the late-glacial and postglacial surface bog deposits that record the northern Europe standard pollen sequence of oldest Dryas, Bølling, older Dryas, Alleröd, younger Dryas, and the younger postglacial pollen zones.

The Netherlands climatic curve combined with a comparable curve derived from Denmark has been considered by some Pleistocene researchers as conclusive proof that the last major period of climatic amelioration preceding the last ma-

jor glaciation must have occurred prior to 50,000 years ago. Of all the chronologies presented in FIGURE 3, the Netherlands curve provides the least continuous and thus the least demonstrable terrestrial chronology available for a sound reconstruction of climate for the time interval 15,000 to 50,000 years ago. The peat layers intercalated in coastal sediments as dated appear, in most cases, to have formed during intervals of more glacial climate as recorded in the other records. It is therefore possible that at least some of these layers represent regressive peat formation during periods of slightly lower sea level, and that the overlying sand and silt layers represent marine transgressions during the recorded intraglacial intervals.

Roosma's Searles Lake pollen curve. The climatic curve derived by pollen analysis of Searles Lake bottom sediments (Roosma, 1958) is shown in FIGURE 3F. Increases in frequency of woodland type pollen at the expense of shrub and herb type pollen are considered to record periods of increasing moisture and decreasing temperature. The pollen curve represents vegetation growing during the last pluvial (Parting Mud) dated between before 21,000 B.C. and 8000 B.C. by Flint and Gale (1958), and during the following period of lake desiccation (Upper Salt). The pollen sequence of the Upper Salt, derived from samples collected at 10-foot intervals does not show the detail of vegetation events shown for the Parting Mud, which is based on samples collected at much closer intervals. The pollen record is consistent with the climatic inferences drawn from the lake-bottom stratigraphy; in addition it provides evidence of secondary climatic fluctuations that seemingly conform in pattern with the dating of the intraglacial intervals ca. 20,500, 17,000, 13,500 and 10,500 B.C. The pollen curve also suggests a consistent pattern of three climatic oscillations between the intraglacial boundaries. As discussed below, the systematic pattern of climatic oscillations suggested by the Searles Lake pollen curve is also reproduced by the more closely dated paleoclimatic records shown in FIGURE 4.

Detailed Chronologies of the Past 25,000 Years

The Cook Inlet glacial curve. In FIGURE 4A the Cook Inlet glacial curve has been replotted to an enlarged time scale to permit emphasis of the secondary glacial oscillations, dated and inferred, of the past 25,000 years. The double vertical lines mark intraglacial culminations, the single vertical lines interstadial culminations.

Michigan basin glacial lake and pollen curves. Zumberge and Potzger (1956) use both geologic and botanical evidence in reconstruction of the hydrologic and climatic history of the Lake Michigan basin. From radiocarbon-dated bluff sections near South Haven, Mich., the sequence of lake-level changes is reconstructed as follows: following high Glenwood Lake a low-water phase (Two Creeks) ca. 9000 B.C.; the Calumet, Toleston and Algonquin (Main Algonquin Lake) shoreline phases between ca. 9000 B.C. and 6000 B.C.; Payette and related lowering lake phases (Late Algonquin Lakes) culminating in low Lake Chippewa between 6000 and 3000 B.C.; transition between low Lake Chippewa and Nipissing Lake phases between 3000 and 2000 B.C.; Nipissing Lake phases between 2000 and 500 B.C.; and Lake Algoma and modern Lake Michigan following 500 B.C. The intercorrelated pollen sequences of the South Haven

buried bog and the nearby Hartford surface bog record an abrupt increase in pine pollen contemporaneous with falling lake levels culminating in the Chippewa low lake, and also an abrupt decrease corresponding to a rise of lake waters to the Nipissing strandlines.

Later papers by Flint (1956) and Hough (1958) are in general agreement with the post-Toleston Michigan Lake basin chronology as worked out by Zumberge and Potzger.* Hough dates the Kirkfield low lake at ca. 7000 B.C. and places it between the Toleston and Main Algonquin strandline phases. New radiocarbon data confirm the placing of the Nipissing high lake stand around 2000 B.C. (Dreimanis, 1958).

The Michigan Lake basin strandline sequence and the pine curves from the South Haven and Hartford bogs as analyzed by Potzger are plotted according to scale in FIGURE 4B. The radiocarbon dated horizons are marked by circles on the curves. The secondary fluctuations of the pollen curves are positioned between the dated horizons on the assumption of uniform depositional rates. The combined lake level and pollen record provides a climatic pattern in close agreement with the Cook Inlet glacial sequence.

Zumberge and Potzger believe that ice retreat successively exposing eastern and lower outlets explains the development of lower strandlines following the main Algonquin level, and that opening of the North Bay outlet culminated in the drop to Lake Chippewa ca. 4000 to 3000 B.C. Radiocarbon dating of glacial deposits in Canada (Karlstrom, 1956; Terasmae and Hughes, 1960), however, shows that the continental ice had receded well north of North Bay and into the Cochrane area much earlier. Thus outlet exposure during continental ice retreat from the Great Lakes cannot be used to explain the lowering lake levels to the lowest Lake Chippewa level, nor an ice advance to explain the following rise to the Nipissing level. The maximum hydrologic and pollen shift in the Great Lakes region coincides with culmination of Naptowne retreat and higher sea levels in Alaska, with the Atlantic period (climatic optimum) in Europe, and with the Altithermal warmest and driest period in western United States. This coincidence therefore suggests a direct climatic explanation for the Chippewa and post-Chippewa lake sequence, that is, warmest and dryest climate during the Chippewa giving way to wetter and cooler climate during the Lake Nipissing and later high-lake phases. Zumberge and Potzger (1956) and Flint (1956), however, place the warmest (thermal maximum) or driest (xerothermic) period of postglacial chronology during Lake Nipissing time on the basis of a few per cent of hickory present in the pollen diagrams.

Northern European climatic curves. Prior to the development of the radiocarbon method for use in dating geologic events, the Swedish varve time scale of DeGeer (1940) covering late-Pleistocene and Recent events comprised the

* By correlations based on assuming similar and contemporaneous strandline levels in the Lake Michigan and Lake Huron basins, Hough considers that lake levels in the Michigan basin reached the Glenwood levels (a 640-foot elevation) during Calumet time. More recently Bretz (1959) marshals convincing data indicating that contemporaneous lake levels in the Michigan basin were lower than those in the Lake Huron basin, and that the Calumet Lake phase was at the 620-foot elevation. According to Bretz the Calumet phase does not represent static lake levels, but two transitory higher levels attained during the general interval of outlet downcutting through glacial drift from the Glenwood level to the stable Toleston bedrock sill level.

only accurately dated geologic chronology available and formed the basis for dating of climatic events inferred from moraines, pollen, and other evidence not only in the Baltic region but, by more or less uncertain correlations, throughout the world. Although the accuracy of the Swedish time scale has been seriously questioned (among others, by Hansen, 1940; Flint, 1947), radiocarbon dating has confirmed the essential accuracy of the Swedish varve chronology. Further confirmation of its accuracy is found in recent remeasurement of varves in Sweden within the time interval 10,000 B.C. to the present (Nilsson, 1960; Järnefors and Fromm, 1960).

Within the integrated late glacial and postglacial climatic chronology of northern Europe, the Danish pollen zones Ia (oldest Dryas), Ib (Bølling), Ic (older Dryas), II (Alleröd), III (younger Dryas), IV (pre-Boreal), V and VI (Boreal), VII (Atlantic), VIII (sub-Boreal), and IX (sub-Atlantic) are directly tied in with glacial retreat and subsequent hydrologic changes in the Baltic Sea area. According to this scheme the middle Swedish moraines (Fennoscandian substage) are correlated with the younger Dryas pollen zone, and the preceding Gotiglacial retreat with the Alleröd pollen zone. Radiocarbon dating of the pollen sequence is in general agreement with these correlations and with the age of the moraines as established by varves; in addition it has resulted in some refinements in subdivision, correlation, and climatic interpretations. The Alleröd has been found to be shorter than previously considered, and the Gothenburg moraines, varve-dated ca. 10,000 B.C., are now considered to have a climatic significance previously unrecognized.

As shown in curve C-1, FIGURE 4, the varve-dated Gothenburg and Middle Swedish moraines (DeGeer, 1957) are contemporaneous with the older Alleröd (now generally referred to as the older Dryas) and with the younger Dryas zones; the preceding and intervening intervals of glacial retreat, respectively, with the Bølling and Alleröd pollen zones as radiocarbon-dated from numerous sites. In curve D-1 of FIGURE 4 the Bølling, older Dryas, Alleröd, younger Dryas, and pre-Boreal climatic zones are represented by the stratigraphy and birch pollen curve of Bølling as worked out by Iversen (1954) in Denmark, the classic region of both the Alleröd and Bølling zones. The dated boundaries shown for the younger Dryas (pollen zone III) represent radiocarbon samples collected from the Bølling section. The older dated boundaries as shown represent average ages of these pollen boundaries dated from numerous other pollen sequences in northern Europe as most recently summarized in Broecker and others (1960).

The postglacial part of the European pollen chronology (curve D-2, in FIGURE 4) is represented by the stratigraphy and birch pollen curve of the closely radiocarbon dated part of the Emmen raised bog in the Netherlands (Zeist, 1958). For direct comparison the birch pollen curve of a radiocarbon dated raised bog in northern Sweden (Lundqvist, 1956) is shown as curve C-2 of FIGURE 4. Although the amplitudes of pollen fluctuations vary appreciably between the two regions, the same general pattern of secondary climatic oscillations may be inferred. Of most significance is the indicated synchronism of the warmest period as inferred from pollen in northern Sweden with the dryest period as recorded in the stratigraphy of the Netherlands raised bog, both culminating ca. 3500

B.C. or precisely with the culmination of the Altithermal (warmest-dryest) as dated in the North American sequence.

Comparison of the radiocarbon dated Scaleby Moss in England with pollen sequences dated elsewhere in Europe leads Godwin and others (1957) to conclude that sufficient radiocarbon data is already available to provide fairly strong evidence for synchronism of pollen zones II, III, and IV as these are defined throughout the North Sea region, and to give some indication of synchronism in the succeeding zones up to Atlantic time (pollen zone VII). Karlstrom (1956) compares the radiocarbon-dated Cook Inlet glacial sequence with the varve-dated northern European climatic chronology and concludes that there is strong evidence for trans-Atlantic synchronism of both small and large climatic oscillations. The new data from Europe appreciably strengthen that conclusion by emphasizing: (1) the climatic significance of a glacial advance ca. 10,000 B.C. (Gothenburg moraines = older Dryas = Mankato maximum = Skilak I); and (2) the probability that as with the North American Altithermal interval, the European Atlantic counterpart was climatically drier, rather than wetter than the preceding (Boreal = Cochrane = Tanya) and succeeding (sub-Boreal = Tustemena) intervals.

The pre-Gothenburg part of the Weichsel (Fourth Glacial) moraine sequence has yet to be definitively dated. The dotted part of curve D-1 in FIGURE 4, is Gross's (1958) hypothetical temperature curve derived from a consideration of radiocarbon-dated climatic data throughout Europe. The dating of a probable interglacial peat at Ale, northern Sweden, at 27,000 B.C. or older (Fromm, 1960) is not incompatible with the interpretation that in Sweden, as in Alaska, a major glacial recession culminating around 45,000 B.C. preceded the last major glaciation (Weichsel = Naptowne). According to the European convention all post-Eem glacial recessions are considered interstadial in rank. The geologic evidence presented in this paper strongly suggests that this pre-Weichsel, post-Eem recession interval, as marked by higher sea levels and major warming of ocean waters, was of interglacial rank, although apparently somewhat cooler than the preceding interglacial (Eem = Sangamon).

Bonneville and Lahontan basin pluvial curves. Radiocarbon dating of strandline tufa and bottom sediments in the Lahontan and Bonneville basins indicates a more complicated series of lake level changes than heretofore proposed (Broecker and Orr, 1958; Feth and Rubin, 1957). Broecker and Orr assess the possibility of errors in dating fresh-water carbonate material and conclude that their dates for carbonate material are probably good to 400 years but may in some cases represent only minimum estimates.

The available radiocarbon data listed in TABLE 2 is used to reconstruct the fluctuations in lake levels shown in curves F-1 and F-2 in FIGURE 4. The tufa and oölite samples are considered to have formed at or just below water levels; the marl and shell samples in deeper water; and the peat, wood, and cultural debris samples in caves and bottom sediments generally above water levels. Thus plotting of these different types according to a time scale and at the elevation from which they were collected provides a curve recording past changes in lake levels.

Periods of low lake levels or desiccation are recorded around 10,500 B.C.

TABLE 2

BONNEVILLE AND LAHONTAN BASIN RADIOCARBON DATES*

Bonneville Basin

Number	Age (B.C.)	Type	Elev. (ft.)
W-385	5770 ± 300	shell	4290
W-335	6380 ± 300	wood	4290
W-386	7780 ± 350	plant	4235
W-326	7975 ± 300	peat	4650
W-440	8310 ± 300	peat	4650
L-435c	8650 ± 300	tufa	4800
L-363c	8950 ± 400	tufa	4800
L-363d	9050 ± 600	tufa	4800
L-435g	9350 ± 250	tufa	4800
W-409	9350 ± 300	tufa	5200
W-439	9470 ± 300	tufa	
W-458	9700 ± 450	tufa	4800
L-363c	10,950 ± 180	tufa	4525
L-382	11,010 ± 350	marl	4650
W-538	11,100 ± 400	tufa	5000–5100
L-435c	11,150 ± 250	tufa	4800
L-363b	11,250 ± 300	tufa	4800
W-491	11,430 ± 400	tufa	4800
W-411	12,050 ± 450	tufa	4520
W-412	12,050 ± 450	tufa	4520
W-445	12,080 ± 500	tufa	4975
W-458	12,430 ± 500	tufa	4800
L-435d	12,850 ± 600	tufa	4800
L-333c	13,250 ± 400	tufa	4690
L-363e	13,580 ± 280	tufa	4800
L-363g	14,150 ± 350	tufa	4800
L-435k	14,450 ± 400	cement	4800
W-490	14,580 ± 800	tufa	4520
W-494	16,050 ± 1000	tufa	4500
W-807	16,550 ± 600	marl	4500
W-876	18,650 ± 500	wood	4775
L-363j	19,250 ± 450	marl	5000–5100
L-363h	21,200 ± 1000	tufa	5000–5100
L-363i	21,350 ± 800	marl	5000–5100
L-33a	23,550 ± 1300	tufa	4520
W-808	27,050 ± 1000	marl	4430
L-33b	31,200 ± 4000	tufa	4800
W-875	38,000 ± 2000	shell	4835
W-410	>30,000	tufa	4800

Danger Cave: elevation 4310 feet

Number	Age (B.C.)	Type
C-640	7010 ± 340	rat dung
C-611	7837 ± 630	charcoal
M-204	8320 ± 650	sheep dung
M-202	8320 ± 650	charcoal
M-119	8450 ± 700	twigs
C-610	9201 ± 570	shrub
C-609	9500 ± 600	sheep dung
M-118 *	9050 ± 700	sheep dung

(No. M118: rerun of C-609)

Lahontan Basin

Number	Age (B.C.)	Type	Elev. (ft.)
L-288f	A.D. 850 ± 200	oolites	3870
L-288h	150 ± 200	shell	3860
L-289r	1250 ± 250	shell	3820
L-364ce	6550 ± 200	tufa	3900
L-364aa	7550 ± 200	tufa	4380
L-289g	7750 ± 200	tufa	4330
L-356h	7750 ± 200	tufa	4280
L-356g	8050 ± 220	tufa	4300
L-364da	8750 ± 250	tufa	4280
L-289i	9300 ± 350	tufa	4360
L-289l	9620 ± 250	tufa	4200
L-289m	9750 ± 200	tufa	4330
L-289c	9750 ± 500	tufa	4005
L-289n	9850 ± 200	tufa	4380
W-442	10,100 ± 400	tufa	4330
L-289h	10,750 ± 300	tufa	3820
L-364am	10,750 ± 300	tufa	4020
L-289s	10,950 ± 350	tufa	4002
L-364an	11,750 ± 300	tufa	4010
L-364cl	12,550 ± 400	tufa	3900
L-289d	12,850 ± 500	tufa	4050
L-289p	13,180 ± 550	shell	4050
L-289aa	13,180 ± 400	tufa	4100
L-289o	13,720 ± 700	shell	4050
L-289k	14,130 ± 750	tufa	4100
L-364cr	14,850 ± 600	marl	4150
L-364cs	15,500 ± 600	shell	4150
L-364al	15,650 ± 650	marl	4005
L-364br	16,750 ± 700	shell	4170
L-364bs	17,800 ± 650	marl	4170
L-289j	26,950 ± 1400	tufa	3950

Fishbone Cave: elevation 3990 feet

Number	Age (B.C.)	Type
L-245	9250 ± 250	twigs
C-599†	9250 ± 570	guano

* Washington radiocarbon laboratory samples (W-numbers) listed in Rubin and Alexander, 1958, 1960; Lamont laboratory samples (L-numbers) in Broecker and Kulp, 1957, Broecker and Orr, 1958, and Broecker, Ewing and Heezen, 1960; Danger Cave dates (C and M numbers) in Jennings (1957).

† Libby, 1955, p. 118.

(L-363-c; L-289-c); 9200 B.C. (L-245, C-599, C-610); ca. 8000 B.C. (W-386, W-326, W-440, M-202); and ca. 7000 B.C. (C-640). Broecker and others (1960) recognize periods of desiccation about 10,500 B.C., 9200 B.C. and a rapid drop from high lake levels after 7500 B.C., but not the desiccation period shown around 8000 B.C. This low lake interval is marked in the Bonneville basin by sample W-386 (plant stems in growth position collected in bottom sediments at an elevation of 4235 feet), samples W-326 and W-440 (peat intercalated in lake sediments at an elevation of 4650 feet) and sample M-202 (charcoal in fire pit of human occupation layer overlain and underlain by sand containing lacustrine ostracods in Danger Cave, elevation about 4310 ft.). These samples representing subaerial environments at low elevation date intermediate to dated tufa samples collected from higher strandline levels and thus must record a period in which lake levels dropped to or below an elevation of 4,235 feet before rising again to an elevation above 4,650 feet as recorded by lake marl overlying peat at that elevation. Broecker considers that the 4,650-foot elevation peat provides excellent evidence for a post-9000 B.C. high-lake stand, and dates new material collected from the site at 7550 B.C. ± 200 and 7950 B.C. ± 200 years. Although the new results provide ages slightly younger than the peat samples analyzed in the Washington laboratory, they are also consistent with the interpretation of a period of desiccation culminating about 8000 B.C. and interrupted by waters rising above the 4650-foot elevation sometime prior to 7500 B.C.

Viewed in the light of the radiocarbon data presented above, Danger Cave provides a strikingly complete stratigraphic record of the late Pleistocene pluvial oscillations. The stratigraphy, as described by Jennings (1957) is briefly as follows, from bottom to top: (1) wave-sorted cemented gravel, more than 6 feet; (2) 2 to 8 inches of cemented sand containing lacustrine ostracod shells (Sand I). Uncharred stem of large shrub near base of sand dates 9200 B.C. ± 570 years (C-610). Charred sheep dung in body of sand dates 8320 B.C. ± 650 years (M-204); (3) human occupation layer marked by remains of small fires built on the surface of Sand I and associated with a few artifacts—charcoal from hearth fire dates 8320 B.C. ± 650 years (M-202); (4) 8 to 12 inches of partly cemented sand containing lacustrine shells similar to those in Sand I (Sand II). Mixed in with the sand are large quantities of sheep dung, charcoal, guano, and twigs, and sticks of wood. Samples collected from Sand II date 9500 B.C. (C-609 sheep dung); 8450 B.C. ± 700 years (M-118-twigs); 7837 B.C. ± 630 years (C-611-charcoal). Rerun of sample C-609 gave a date of 9050 B.C. ± 700 years (M-118); (5) Thin discontinuous layer of bat guano that seems to have blanketed the entire cave and marks a short time of bat occupancy of the cave prior to heavy human occupancy as recorded in overlying loessal-type deposits. Charred rat dung associated with the guano layer dates 7010 B.C. ± 340 years (C-640); and (6) 8 to 12 feet of alternating beds of eolian dust and vegetable matter containing cultural debris and separated into four distinct zones by thin discontinuous layers of angular limestone fragments (talus and roof-fall material). Twigs collected from the loessal material date between 2850 B.C. ± 350 years and A.D. 20 ± 240 years (samples M-205, M-203, C-636, and C-635).

Jennings believes it probable that the cave was not occupied during the

depositional periods of Sand I and Sand II, and thus that the organic material in the sands was derived from overlying and underlying culture layers as a result of incorporation by churning and mixing during and after deposition of the thin sand beds. Within the limits of dating error the age of the samples collected from within the sand units is consistent—with the possible exception of C-609—with the interpretation that they represent material incorporated from underlying and overlying culture layers. The maximum and minimum dates for Sand I are 9200 B.C. (C-610) and 8320 B.C. (M-202) and, for Sand II, 8320 B.C. and 7010 B.C. (C-640). Thus, as dated, Sand I was deposited when lake waters occupied the Provo level ca. 8500 B.C., and Sand II was deposited when the Provo level was reoccupied ca. 7500 B.C.

This interpretation differs from that of Eardley and others (1957, p. 1169), who consider that the basal gravel unit represents the last time Danger Cave was flooded by waters of Lake Bonneville, and from Jennings (1957, p. 60 and 91), who believes that Sand I represents the last time the cave was below lake water levels and that Sand II is a windblown deposit.* The major stratigraphic break in the cave sequence is marked by the abrupt change from sand deposition to loessal and cultural debris deposition beginning ca. 7000 B.C. and continuing to the present, and the most reasonable interpretation is that it records the last time lake waters receded below the cave mouth.

From detailed stratigraphic studies of bottom sediments in the Lahontan basin, Morrison (1952) works out a sequence of three Lahontan high-level phases and seven post-Lahontan lower lake-level phases. Sediments of the three Lahontan phases are separated by a disconformity and a peat layer. A weak desert soil separates Lahontan III from the first post-Lahontan phase; a stronger desert soil (the Altithermal soil), the first and second post-Lahontan phases; and disconformities separate the sediments of the later post-Lahontan phases. As shown in G, FIGURE 4, this stratigraphic sequence is seemingly compatible with the radiocarbon-dated strandline sequence, and records the same number of post-Altithermal events as derived from the Cook Inlet evidence. Richmond, Morrison and Bissell (1952) provisionally correlate the post-Lahontan I phase with the Cochrane, and the three older Lahontan pluvial phases respectively with the Iowan-Tazewell, Cary, and Mankato substages. The radiocarbon data, as interpreted in this paper, confirm their Cochrane correlation but indicate that the last three higher Lahontan phases are all of Mankato age.

* Jennings believes that the eolian origin of Sand II is indicated by: (1) a dunelike form with maximum thickness near the cave mouth, and progressive thinning of the sand in towards the interior of the cave; (2) the abundance of broken ostracod shells which, he assumes, represents breakage by wind action; and (3) the presence of much woody material in the underlying Sand I, which he believes would have been floated away if Sand II were water-deposited. No mention is made of frosted sand grains, eolian crossbedding, or other primary structures used in distinguishing between windblown and waterlaid deposits. Jennings criteria are not by themselves definitive of an eolian origin. The dunelike form can be as readily explained as a beach ridge or subaqueous bar feature. Buried wood is not an uncommon component of strandline and of other types of waterlaid deposits. It is not clear why wind action would be more effective in breaking shells than wave or current action. Furthermore, Sand II comprises the upper layer of the floor of the cave upon which heavy traffic by man and animals took place, as recorded by the overlying culture layers and the mixed and churned aspect of the sand itself. The probability is thus indicated that much of the shell breakage in Sand II took place after, rather than during, deposition.

Dated tufa and marl samples from both the Lahontan and Bonneville basis are consistent in suggesting lower lake fluctuations or periods of desiccation ca. 11,500 B.C., 12,500 B.C., and 13,500 B.C. Before 13,500 B.C. the data are too sparse to suggest anything but the broadest shifts in lake levels from maximum levels apparently attained around 20,000 B.C.

In support of their climatic theory, Ewing and Donn (1956) use the Bonneville and Lahontan pluvial sequence as additional proof of an abrupt period of warming in the late Pleistocene chronology centered around 9000 B.C. In this they follow Broecker and Orr's (1956, 1958; also Broecker's, 1957) interpretation that the last pluvial phase was terminated abruptly around ca. 9200 B.C. by a period of desiccation that has continued to the present. This interpretation is no longer tenable, as shown in FIGURE 4. The short period of desiccation ca. 9000 B.C. was followed by two later and major pluvial phases. Furthermore, the pluvial evidence is consistent with the other paleoclimatic evidence presented in recording more important periods of warming ca. 10,500 B.C. and 7000 B.C. immediately preceding and following the 9000 B.C. climatic oscillation. Thus to date the end of the Wisconsin glacial stage ca. 9000 B.C. is to place the boundary of a major geologic subdivision at a distinctly subordinate climatic oscillation.

Wiseman's ocean temperature curve. From meticulous analyses, both mechanical and chemical, of samples collected at 2-cm., intervals from a carefully selected deep sea core, Wiseman (1958, 1959) has reconstructed the most detailed and closely dated ocean temperature record known to me. Each sampled section is analyzed for size, grade, carbonate content, and fossils, and then dated by its titanium and iron content. The basic assumption used in dating is that the rates of titanium and iron deposition have been uniform over the time interval of the core record. Wiseman's conclusion that variations in time of carbonate content closely reflect changes in water temperatures (curve I-1, FIGURE 4) is confirmed by oxygen-isotope temperature analyses of numerous duplicate samples (curve I-2, FIGURE 4).

The agreement between the Cook Inlet glacial curve and Wiseman's temperature curves is striking. All the intraglacial retreatal intervals of the Cook Inlet sequence closely coincide in time with major intervals of surface ocean-water warming as recorded in the equatorial core, ca. 20,000 ca. 17,000, ca. 13,500, ca. 10,500, ca. 7000, ca. 3500, and ca. A.D. 500. The main subdivisions of the Skilak, Tanya, and Alaskan substages of the Cook Inlet record coincide within the limits of dating error with subordinate intervals of ocean-water warming in the core. The amplitude of the secondary temperature oscillations diminishes in the lower part of the curve recording the coldest surface water. This may be in part the result of generally greater spacing between dated sections in this part of the curve, but it probably largely reflects dampening of secondary temperature oscillation effects on deposition during the period of maximum glaciation. There is but a suggestion of subordinate temperature fluctuations in the Killey substage part of the deep sea core, and no hint of such fluctuations at all in the pre-Killey part.

Wiseman notes slight discrepancies between his carbonate- and isotope-temperature curves and believes (oral communication, 1960) that the carbonate curve provides the truest record of past climatic change. These discrepancies

are perhaps within the limits of dating error and temperature analysis. However, it may be significant that where such discrepancies exist, the isotope-temperature curve agrees most closely with the Cook Inlet record. Thus, whereas the carbonate curve indicates culminations of sharp warming trends ca. 6600 B.C. and 3000 B.C., the isotope temperature curve dates them at ca. 7000 B.C. and 3500 B.C., or identical with the dating of the Skilak/Tanya and Tanya/Tustemena boundaries in the Cook Inlet sequence. Likewise, whereas the carbonate curve indicates maximum warming ca. 1200 A.D., the isotope temperature curve dates the period of maximum warmth ca. 3500 B.C. or coincident with the major boundary marking maximum glacial retreat and higher sea level used in the Cook Inlet chronology to separate the Naptowne from the Alaskan glaciation.

Considering the great differences in character of the Cook Inlet and equatorial core records (terrestrial glacial versus marine temperatures), different methods used in dating, and the spread of more than 60 degrees latitude and more than 130 degrees longitude between depositional sites, the similarity in paleoclimatic pattern shown is somewhat unexpected and suggests that the prime cause of these secondary climatic oscillations was widespread (if not world-wide) and synchronous in effect, and therefore, almost certainly, extraterrestrial.

Fairbridge's glacioeustatic sea-level curve. By selecting radiocarbon samples reflecting past sea-level changes in stable coastal areas, Fairbridge (1958, 1960) has reconstructed a sea-level curve over the past 10,000 years that he attributes to glacioeustacy. The agreement between his curve, the glacial chronology of Cook Inlet and correlated paleoclimatic records presented above provides strong support for this interpretation. Discrepancies between his curve (as curve E in FIGURE 4, inverted to facilitate comparison with the glacial and pluvial records) and the Cook Inlet glacial curve fall within the limits of error of dating geologic events by radiocarbon techniques.

Detailed Chronologies of the Past 3000 Years

The Cook Inlet glacial curve of Alaskan time. The most detailed part of the Cook Inlet glacial chronology is inferred from dated tidal-bog stratigraphy and moraines. Numerous tidal bogs exposed locally near sea level in Cook Inlet show alternating beds of peat and tidal silt; the peat layers represent times when sea levels were generally lower than at present, and the overlying tidal silt layers represent times of rising sea levels. A particularly complete tidal-bog section is exposed near Girdwood at the head of Turnagain Arm. Here a 15-foot section is exposed during extreme low tides and comprises 26 definable stratigraphic units of woody peat, sedge peat, mixed silt and peat, and tidal silt (FIGURE 5). The lowest exposed peat bed (*a*) dates near its upper contact 850 B.C. ± 180 yrs. (W-299); a higher woody peat layer (*r*) dates A.D. 1250 ± 250 (W-175); and tree ring counts on a stump rooted in the uppermost woody peat layer (*y*) indicates an age of about A.D. 1850 for this layer. The surface layer in the bog is of mixed peat and silt suggesting progressive rise in sea levels since ca. 1850 A.D. Yearly observations in the Girdwood area since 1950 revealed progressively thicker silt accumulations on the bog surface beginning in 1954 and indicating that high spring tide depositional levels have risen once again to above the general level of the bog surface along its coastal margin.

Cross dating of this tidal-bog sequence with the morainal sequence in the nearby Kenai Mountains indicates that the lower sea level periods, marked by regressive peats, coincide respectively in time with glacial advances to the Tustemena III, Tunnel I, and Tunnel II end moraines, and that the higher sea-level periods, marked by transgressive tidal silt layers *b*, *c*, *j*, *l*, and *s* coincide with retreatal culminations immediately preceding these glacial advances. Thus the combined record strongly suggests glacioeustatic sea-level shifts in keeping with the geoclimatic equations: glacial advance = lowering sea level and peat deposition; and glacial retreat = rising sea level and tidal silt deposition. As reconstructed in FIGURE 5, the inferred glacial curve is time-calibrated according to most probable age of the analyzed C-14 sample (no manipulation of the $\pm$ of any sample has been made) and the intermediate oscillations between dated boundaries are drawn as a first approximation on the assumption permitted by stratigraphic relations that they represent events of approximately equal duration. The possible errors involved in this inductive approach are considered to be less than those that would result from adjusting the curve according to a preconceived notion of what the climatic history of the time interval involved should be.

A glacioeustatic sea-level chronology records relative changes in the total volume of the world's land ice. Thus the agreement between sea-level shifts and glacial oscillations in Cook Inlet by itself strongly suggests that the glacial record of this Alaska region may be considered representative of average climatic changes throughout the world. That this is generally true over a much longer time interval is supported by the agreements between the dated Cook Inlet glacial sequence and the other independently dated paleoclimatic records discussed above and shown in FIGURES 3 and 4. However, direct testing of the degree of representativeness of the subordinate oscillations, as dated in the tidal bog sequence, with the climatic history of other regions is handicapped by the general absence of precisely dated, detailed, and continuous primary depositional sequences from other regions. In FIGURE 5, the reconstructed glacial curve of Cook Inlet is compared with the historical Nile River record of flood-level fluctuations, as discussed by C. E. P. Brooks (1949); the detailed recurrence surface and pollen sequence worked out for the Sпöromsmossen bog near Stockholm, Sweden, by Ording and von Post (von Post, 1946); a tree-ring curve that I obtained in 1954 from cores taken from 5 mountain hemlock trees located generally above the timberline on the Cook Inlet side of the Kenai Mountains; Schulman's (1956) upper timberline limber pine curve of the northern Rocky Mountains; part of the C-14 dated glacioeustatic sea-level curve of Fairbridge (1958, 1960), and archaeologically dated marine transgressions of coastal Netherlands (Bennema, 1953).

Historically recorded Nile River flood fluctuations. According to Brooks (1949), the most important source of information on rainfall variations in Africa is provided by the historically recorded fluctuations in floods of the Nile River, both at low- and high-flood levels. The high flood-level fluctuations are interpreted as primarily representing summer and early fall monsoonal rainfall variations in Abyssinia, the headwater area of the Blue Nile, the Atbara, and the Sobat Rivers between 5° and 10° N lat. The low flood-level fluctuations are considered to reflect primarily spring rainfall in the equatorial belt of low

pressure. Both the low- and high-flood fluctuations for the recorded interval since A.D. 640 show a general upward rise at a rate of about 10 cm. per century. Brooks attributes this rise to progressive sedimentation in the lower Nile. However, sedimentation levels at the mouths of rivers are primarily determined by the level of sea level, and the secular rise of 10 cm. per century is of the same general order of magnitude as that of the glacioeustatic sea-level rise described from other coastal areas of the world (Fairbridge, 1960). Thus there is strong geologic evidence that the Nile low-flood level fluctuations may provide a sensitive index of sea-level fluctuations in the Mediterranean Sea rather than simply a regional spring rainfall index as interpreted by Brooks. If so, then the low flood-level fluctuations (sea level) could be expected to provide a curve that is reciprocal to the high flood-level curve (precipitation). When corrected to constant low flood levels, the high flood levels provide a curve (FIGURE 5) which is a reciprocal of the low flood-level curve. The resulting precipitation index of the 5° to 10° N lat. region parallels the Alaskan glacial record as dated, with maximum periods of precipitation in this part of Africa coinciding with the glacial maxima dated ca. A.D. 800, ca. 1000, and ca. A.D. 1300; and with periods of minimum precipitation coinciding with glacial retreats dated ca. A.D. 850, A.D. 1150, and A.D. 1450. The post-A.D. 1450 part of the Nile record is distorted by construction of an irrigation system in the delta region and is therefore not useful for the purpose of interpreting past climatic change.

Snöromsmossen recurrence surface and pollen record. As previously discussed (Karlstrom, 1956) the recurrence surface sequence of Sweden as dated by Granlund (1932; Magnusson and others, 1947) from archaeological, pollen, and varve evidence, records a series of wet and dry periods of peat accumulation in general climatic agreement with the radiocarbon-dated Cook Inlet glacial record. Detailed work on the bog and pollen sequence from Snöromsmossen near Stockholm, one of the chief localities of Granlund's investigations of recurrence surfaces and their dating, reveals the presence of more weakly developed recurrence surfaces between those described by Granlund (von Post, 1946). Ording's meticulous pollen analysis of the bog, based on sampling at 2-cm. intervals, shows that the absolute pollen amount regularly rises and sinks with the lower and higher degree of humification of the peat, and that the birch pollen frequency characteristically decreases during the intervals of peat humification and recurrence surface development. Von Post analyzes the birch curve and finds by progressive smoothing that it is composed of 5 elementary curves, 4 of which are clearly rhythmic, and shows a direct relationship to the recurrence surfaces. The fifth and most generalized curve shows the ordinary diminution of woodland birch towards the present day. Von Post concludes that these 5 curves, smoothed respectively by joining the mean figures for 2, 4, 7, 14, and 30 analyses values, record true climatic shifts in the past, since these are registered consonantly by 2 reactions wholely independent of each other: (1) the local depositional conditions recorded by the alternation of dry and wet periods of bog accumulation; and (2) shifts in the composition of surrounding forests as recorded by the pollen frequencies.

The Snöromsmossen record suggests the same climatic pattern derived from the Cook Inlet evidence and, within the limits of dating error, it can be directly

correlated with the glacial sequence (FIGURE 5). Curve 1 represents the birch pollen curve smoothed by averaging 2 pollen analyses; curve 2 averages 7 analyses. The recorded retreatal oscillations in Alaska are matched in each instance by a recurrence surface in the Snöromsmossen bog contemporaneous with a decreased birch pollen frequency (curve 1). In addition the Snöromsmossen bog records 2 additional weakly developed recurrence surfaces ca. A.D. 450 and A.D. 1550 of approximately one half the wave lengths of the other recorded oscillations. The more generalized birch curve 2 emphasizes periods of maximum dryness ca. 500 B.C., ca. A.D. 500, and ca. A.D. 1500 in Sweden, or coincident with major retreatal phases just preceding the Tustumena III, Tunnel I, and Tunnel II advances. The lack of precision of dating by radiocarbon and by archaeological methods obscures any regional climatic or depositional lags that may conceivably be represented in the secondary oscillations of the Alaskan and Swedish records.

Archaeologically dated marine transgressions in the Netherlands. Bennema (1953) summarizes archaeological data relating to past marine transgressions in coastal Netherlands. Transgressions are separated by regressions or static conditions, and are dated roughly at 700–300 B.C., about the beginning of the Christian era, A.D. 300–600, during the latter part of the 9th and during the 10th century, intermittently during the 14th, 15th, and 16th centuries, and following A.D. 1850. Bennema correlates this transgressive series with the Scandinavian recurrence surface sequence and with other climatic evidence, and concludes: (1) that the transgression sequence is probably climatic; and (2) that it suggests a periodicity of about 525 years with a tendency for every other transgressive phase to be of larger magnitude. Considering the difficulties of defining climatic transgressive phases in a coastal region that is known to have been subject to both isostatic uplift and subsidence, and the uncertainties involved in dating from scattered archaeological evidence, the Netherlands record as interpreted by Bennema is in reasonably good agreement with the more closely stratigraphically defined Cook Inlet sequence (FIGURE 5). The fact that glacial fluctuations in Cook Inlet were seemingly synchronous with the minor sea-level oscillations recorded from the Mediterranean Sea (the Nile record), the North Sea (the Netherlands record), and the north Pacific (the Cook Inlet tidal bog record) substantially strengthens Bennema's interpretation that the Netherlands transgressions represent glacioeustatic or climatic events.

Glacioeustatic sea-level curve. Fairbridge's (1960) sea level curve reconstructed from radiocarbon-dated samples collected from stable coastal regions in a broad fashion reproduces the same pattern of sea-level shifts inferred from the evidence presented above (FIGURE 5). Fairbridge (1958) notes two areas of potential error in the reconstruction of his curve: (1) dating uncertainties associated with C-14 analyzed samples, particularly those derived from analyses of carbonate material; and (2) the difficulty of determining the past levels of sea level from the ecology, stratigraphic position, and elevation of the dated samples. Thus the discrepancies between his curve and the Cook Inlet and correlated evidence may be more apparent than real. Fairbridge's curve shows sea levels higher than present datum ca. A.D. 850 and ca. A.D. 1100. However, the Cook Inlet, Nile River, and perhaps the Netherlands evidence as well, suggest that sea level remained below present datum during these intervals.

Fairbridge's curve shows a regressive phase around 1500 A.D. This is contrary to the evidence of higher sea levels in Cook Inlet, the North Sea, the Mediterranean Sea, and with the evidence of climatic dryness in Sweden ca. A.D. 1500.

Tree-ring chronologies. Major efforts have been made to infer past climatic changes from shifts in the annual growth rate of trees. Conditions controlling the growth of trees are complex and include many factors other than climate. It is particularly important to sample trees from environments in which climate is most likely the dominant threshold factor affecting growth rates. Schulman (1956, p. 9) concludes that the numerous ring chronologies of rainfall or temperature presently available show that both the drought chronologies in the warm dry lands and the temperature chronologies in the north lands tend to reach greater sensitivities as the sites become more limiting in terms of the respective climatic element. The two tree-ring chronologies selected for consideration in this paper are representative of extreme growth environments associated with the uppermost tree lines in a cold subarctic alpine region, and in the semiarid northern Rocky Mountain region.

The mean annual indices of the 5 mountain hemlock trees above timberline that I sampled in 1954 are given in TABLE 3. The smoothed curve (FIGURE 5) is based on running means centered at quarter-century intervals. Historically recorded glacial advances between A.D. 1600 and 1620, the beginning and middle of the 18th century, and the middle of the 19th century generally coincide with periods in which the timberline hemlocks were experiencing progressively more difficult growth conditions. Comparisons with the limber pine curve of the northern Rocky Mountains, comparably smoothed from annual indices supplied by Schulman (1956, p. 77 and 80), reveal a strong negative correlation. The periods of poor tree growth in Alaska (low temperature) coincide with the periods of optimal growth (high precipitation) in the timberline region of the northern Rocky Mountains.

In so far as periods of glacial advance are favored by both wet and cold climate, the combined tree-ring record is consistent in suggesting climatic shifts in the northern Rocky Mountains parallel with those in Alaska. To facilitate comparison between the two tree-ring records and the Cook Inlet glacial curve, the Alaska mountain hemlock curve (thermograph) is inverted. Significant drought and warming trends culminating near the beginning of the 18th and 19th centuries occurred during intervals of higher sea level and glacial retreat in Tunnel II time. The maximum advance of Tunnel II ice coincides with low temperatures and increased precipitation as shown by the two tree-ring curves. The longer limber pine curve shows additional areas of agreement with the Cook Inlet glacial curve. The major shift towards low precipitation is centered around A.D. 1500, or contemporaneous with the important Tunnel I/II retreatal boundary, and lesser but important intervals of lower precipitation reflected in the limber pine curve line up fairly well with retreatal intervals recorded during Tunnel I time.

Comparison with Schove's sunspot curve and the atmospheric radiocarbon variations curve. Variations in sunspot numbers have been directly correlated with glacial and sea level fluctuations (Thoraninsson, 1940; Lawrence, 1950; Flint,

1953), with tree-ring growth (Douglass, 1936), and with changes in world weather and geomagnetic patterns (among others, Willett, 1943). Most recently Fairbridge (1960) correlates his glacioeustatic sea-level curve with

TABLE 3

TREE-RING INDICES (MILLIMETERS), KENAI PENINSULA
TIMBERLINE MOUNTAIN HEMLOCK; FIVE TREES

A.D.	0	1	2	3	4	5	6	7	8	9
1570										0.39
1580	0.14	0.18	0.21	0.28	0.11	0.25	0.28	0.35	0.18	0.28
1590	0.25	0.39	0.35	0.53	0.32	0.28	0.42	0.09	0.56	0.11
1600	0.18	0.18	0.14	0.28	0.28	0.35	0.28	0.07	0.09	0.07
1610	0.11	0.25	0.18	0.21	0.25	0.05	0.04	0.09	0.11	0.14
1620	0.07	0.11	0.04	0.14	0.11	0.16	0.14	0.07	0.14	0.09
1630	0.18	0.14	0.14	0.14	0.14	0.35	0.35	0.32	0.28	0.44
1640	0.44	0.39	0.39	0.33	0.30	0.53	0.44	0.39	0.35	0.39
1650	0.33	0.25	0.25	0.33	0.25	0.14	0.21	0.21	0.23	0.23
1660	0.16	0.19	0.18	0.32	0.23	0.25	0.21	0.13	0.09	0.12
1670	0.16	0.14	0.09	0.13	0.19	0.19	0.19	0.14	0.18	0.18
1680	0.18	0.21	0.21	0.16	0.16	0.15	0.14	0.08	0.05	0.06
1690	0.08	0.11	0.11	0.18	0.18	0.16	0.14	0.23	0.23	0.16
1700	0.21	0.26	0.30	0.21	0.30	0.23	0.14	0.17	0.22	0.14
1710	0.15	0.18	0.18	0.16	0.14	0.14	0.18	0.09	0.14	0.19
1720	0.15	0.21	0.53	0.43	0.51	0.50	0.69	0.40	0.51	0.55
1730	0.55	0.61	0.45	0.37	0.34	0.26	0.33	0.32	0.34	0.32
1740	0.43	0.46	0.43	0.49	0.44	0.34	0.34	0.36	0.34	0.26
1750	0.31	0.26	0.26	0.38	0.40	0.32	0.35	0.29	0.41	0.38
1760	0.32	0.32	0.25	0.46	0.47	0.57	0.49	0.42	0.37	0.36
1770	0.32	0.32	0.33	0.36	0.39	0.39	0.35	0.43	0.36	0.39
1780	0.32	0.41	0.39	0.35	0.43	0.64	0.63	0.64	0.63	0.60
1790	0.62	0.59	0.61	0.57	0.57	0.52	0.56	0.74	0.56	0.59
1800	0.66	0.60	0.70	0.65	0.67	0.78	0.82	0.86	0.85	0.62
1810	0.70	0.59	0.38	0.33	0.39	0.38	0.33	0.30	0.34	0.33
1820	0.35	0.37	0.39	0.37	0.41	0.44	0.41	0.46	0.48	0.61
1830	0.60	0.72	0.65	0.70	0.57	0.53	0.55	0.52	0.46	0.54
1840	0.56	0.67	0.72	0.67	0.62	0.56	0.45	0.56	0.49	0.50
1850	0.42	0.50	0.42	0.45	0.41	0.50	0.36	0.36	0.36	0.43
1860	0.39	0.41	0.42	0.42	0.43	0.45	0.50	0.56	0.68	0.60
1870	0.63	0.56	0.52	0.65	0.67	0.65	0.71	0.62	0.54	0.48
1880	0.39	0.31	0.25	0.26	0.26	0.33	0.30	0.41	0.38	0.41
1890	0.41	0.46	0.44	0.46	0.45	0.51	0.53	0.44	0.39	0.36
1900	0.41	0.43	0.45	0.50	0.48	0.44	0.53	0.53	0.61	0.54
1910	0.59	0.39	0.36	0.33	0.36	0.34	0.38	0.39	0.43	0.41
1920	0.45	0.40	0.46	0.49	0.50	0.64	0.59	0.65	0.55	0.51
1930	0.55	0.57	0.56	0.54	0.41	0.54	0.53	0.50	0.44	0.43
1940	0.50	0.50	0.53	0.60	0.48	0.51	0.46	0.62	0.54	0.62
1950	0.40	0.33	0.32	0.40	0.43					

* Collected by T. N. V. Karlstrom, 1954; measured by Elizabeth B. Coulter, 1955.

Schove's (1955) sunspot curve of 30-year smoothed sunspot maxima. The smoothed sunspot curve is reproduced in FIGURE 5.

The curve of atmospheric radiocarbon variations over the past 1300 years (Willis and others, 1960) is also reproduced in FIGURE 5. This curve is derived by analysis of the radiocarbon activity of individual rings of a sequoia tree sampled at 50-year intervals. Conversion to percentage variations in C-14 activity of each ring is made relative to the analyzed activity of the A.D. 1859

ring. The check analyses run by three different radiocarbon laboratories resulted in some anomalous measurements. However, the general trend of the curve is considered significant by the authors, and they believe it confirms the existence of short term oscillations in radiocarbon concentration, perhaps with a period in the order of 150 to 200 years, superposed on an oscillation having a longer period of the order of 1200 years.

There is general agreement between this curve and the radiocarbon concentration curve obtained for the last 300 years by DeVries and others (1958). DeVries correlates the radiocarbon variations with the historically recorded advances and retreats of glaciers. This suggests a link between variations in atmospheric radiocarbon and climatic phenomena. Willis and others suggest that the radiocarbon variations could have resulted either from variations in the cosmic ray flux, perhaps in association with changes in the earth's magnetic field or from fluctuations in the isotope equilibrium of the carbon-exchange reservoir. M. Stuiver (unpublished) suggests the probability of a negative correlation between sunspots and observed variations in atmospheric radiocarbon concentrations.

As shown in FIGURE 5, comparison of the sunspot and atmospheric radiocarbon variation curves with the paleoclimatic record is handicapped by the variability of oscillations shown by the two types of record. No simple correlation is evident or possible. Nevertheless some broad relations are indicated that bear directly on possible physical mechanisms linking solar activity with terrestrial climate. The recorded intervals of glacial retreat tend to coincide with intervals of low sunspot activity and of high atmospheric radiocarbon concentrations, whereas the intervals of glacial advance tend to be associated with periods of higher sunspot activity and low atmospheric radiocarbon concentration.

The apparent negative correlation between radiocarbon atmospheric concentrations and sunspot numbers emphasizes the fact, generally recognized, that relative sunspot numbers provide no direct or simple index of solar radiation intensity changes. The apparent correlation between periods of glacial retreat and periods of low sunspot maxima (particularly well shown around A.D. 1450 and 1700) is contrary to Willett's (1953, p. 64) generalization that the periods of low sunspot activity have tended definitely toward the glacial in climatic pattern, and those of high activity toward the interglacial. Although Willett's meteorological analysis remains one of the most cogent statements in support of the solar emission climatic theory, the paleoclimatic data presented here suggest that some of the assumptions and premises used in the development of the theory need critical review. As will be discussed below, it is possible that some of the geophysical and climatic phenomena attributed to variations in solar emission may more directly reflect geometric variations in solar insolation and tidal forces resulting from planetary movements about the sun.

Comparison of the Naptowne and Alaskan paleoclimatic sequence with the Pettersson tide-generating force curve. Pettersson (1914) proposed a theory correlating changes in tidal-force intensity with what he referred to as long-range climatic variations. From a consideration of the inclinations and proximity of the earth to moon and sun, calculations based on celestial mechanics indicate that intervals of increased tidal intensity occur periodically on the

average of every 1700 years and that subordinate cycles of about 90 and 9 years are superposed on the larger periodicity. According to Pettersson's calculations, tidal intensity maxima occurred around A.D. 1433, 250 B.C., 1800 B.C., 3500 B.C., and before then on an average of every 1700 years. These periods of tidal maxima coincide with times when the sun, earth, and moon are in line with earth at perihelion and the moon at perigee.

From the paleoclimatic evidence available to me in 1954, I concluded that the emerging climatic pattern strongly suggested widespread if not global climatic synthronism in the past and a systematic series of superposed cycles, including cycles of ca. 550, 1100, 3400 and 40,000 years, seemingly interrelated in a harmonic or near-harmonic fashion with each other and with the ca. 40,800-year astronomic obliquity cycle and the ca. 1700-year Pettersson tide-generating force cycle (Karlstrom, 1955). As shown in FIGURES 2, 3, 4, and 5, the new paleoclimatic evidence would appear to strengthen the case both for the climatic synchronism and for climatic cyclicity.

The degree of parallelism between secondary climatic oscillations recorded by the various types of paleoclimatic sequence may best be judged by a review of the evidence presented in FIGURES 3 and 4. The vertical lines shown on these plates mark the culmination of important retreatal (warming) phases as these are dated and inferred from the Cook Inlet chronology. It is to be emphasized that the other paleoclimatic curves have not been arbitrarily adjusted or fitted to these Cook Inlet boundaries on the conventional assumption of parallel sequence. Only those records have been included that are dated closely enough from internal evidence to permit direct and independent plotting according to a common time scale. To minimize further the possible subjective tendency towards curve fitting, the radiocarbon-dated curves are positioned according to the most probable age of the samples as reported by the laboratories or interpreted in the literature; no manipulation has been made within the quoted range of dating error except in one case where such an adjustment is required by stratigraphic relations.* Thus the consistent in-phase relationships shown by the various curves would appear not only to support their validity as climatic indicators, but also the general accuracy of the various dating methods used in their reconstruction.

In FIGURE 6 the composite paleoclimatic record is compared with the Pettersson tide-generating force curve and the Milankovitch solar-radiation curve. The solid vertical lines mark culminations of interglacials, intraglacials, and interstadials that are recorded and directly dated in one or more of the paleoclimatic sequences previously discussed. The dashed vertical lines represent comparable boundaries inferred from the total paleoclimatic pattern but for which there is as yet no direct record.

The apparent relationship between stage rank events and the Milankovitch curve has been discussed previously. A much more direct relationship is suggested between the reconstructed secondary paleoclimatic oscillations and the Pettersson tide-generating force curve. The intraglacial boundaries dated or

* In Brandtner's east-central European climatic curve, one radiocarbon sample dates seemingly older than a sample collected from a lower stratigraphic horizon. Within the limits of counting error the higher sample can be considered younger than the lower sample. This adjustment, within the limits of quoted dating error, is made in curve G, plate 3.

inferred to the nearest 500 years at ca. 3500, 7000, 10,500, 13,500, 17,000, 20,500, 24,000, 27,500, 31,000, 34,500, 38,000, 41,500, and 45,000 B.C., closely coincide with alternate tidal intensity maxima dated ca. 3500, 6900, 10,300, 13,700, 17,100, 20,500, 23,900, 27,300, 30,700, 34,100, 37,500, 40,900, and 44,300 B.C. by celestial mechanics. This coincidence appears to be too close to be accidental.

The intermediate tidal intensity maxima have no apparent record in the paleoclimatic sequences except in the most sensitive ones as shown in FIGURE 5, where retreatal culminations reflecting the ca. 550-year cycle coincide with both the 250 B.C. and A.D. 1433 tidal intensity maxima. In the other detailed but less sensitive paleoclimatic curves the major subdivision of the substage oscillations is apparently determined by the ca. 1100-year cycle. The climatic or astronomic reason for more pronounced climatic effects to be associated with alternate tidal-intensity maxima is not clear,* but it may have a harmonic solution. Superposition of a ca. 1100-year cycle on the 1700-year tidal cycle would result in out-of-phase relations at alternate maxima and result in a reinforced effect at twice the 1700-year cycle to provide the substage cycle of ca. 3400 years. The persistance of the ca. 3400-year substage cycle and its consistent threefold subdivision as independently recorded and dated in the diverse types of paleoclimatic sequence over a time range of more than 15,000 years supports the reality of this climatic pattern and its probable harmonic nature.

The data presented in FIGURE 5 extend the impression of a direct harmonic relation between terrestrial climate and the tide-generating force curve. The pattern of climatic oscillations seemingly reflects the subharmonic cycles of ca. 1100, 550, and 280 years directly in phase with the tidal-intensity maxima dated ca. 250 B.C. and A.D. 1433. In this time interval the Pettersson cycle has a calculated wave length of 1683 years; its subharmonics as schematically plotted in FIGURE 5 have wave lengths thus of 561, 280.5, and 93.5 years. The vertical lines on FIGURE 5 are not drawn directly from the paleoclimatic evidence as in FIGURES 2, 3, and 4, but are positioned on the tidal-intensity maxima of the theoretical tidal force curve.

The $\frac{1}{18}$ subharmonic (shown as dashed lines in FIGURE 5) is the equivalent of Pettersson's ca. 90-year subordinate tidal cycle: it has no apparent reflection in the paleoclimatic curves shown. However, it may be significant that the $\frac{1}{18}$-subharmonic curve in a broad way reproduces the major oscillations of the smoothed sunspot curve, with tidal-force maxima tending to coincide with periods of lower sunspot numbers and thus also with periods of higher radiocarbon atmospheric concentrations. This would suggest some indirect physical linkage between these various geophysical phenomena and introduces the possibility that it may eventually prove possible to integrate sunspot, atmospheric radiocarbon concentration, and other geophysical phenomena related to terrestrial climate into one comprehensive geometric variations theory.

C. E. P. Brooks (1949, pp. 368 to 373) accepts Pettersson's tidal-force curve as an important but little understood factor in the climatic trends of the past

* C. M. Stacey (unpublished) calculates the "maximum perigee spring tides" (the Pettersson cycle) at ca. 1668 years. He notes (written communication, February, 1961) that the moon's perigee seems to fall on opposite sides of the earth relative to the sun at the close of alternate maxima; also that the lunar nodal and eclipse cycles are in phase only at alternate maxima.

5000 years. Both he and Pettersson assume that the tidal force shifts affected climate indirectly by determining changes in ocean circulation. It is further assumed that the climatic effect of the tidal forces are restricted to the past 5000 years because appropriate sea-level and submarine threshold conditions did not exist prior to that time.

More recently it has been argued that ocean circulation as a factor in climate must be relegated to a role distinctly secondary to changing patterns of atmospheric circulation. As previously discussed, Willett (1953) proposes that the primary cause of the atmospheric circulation changes is the type of solar emissive activity of which such phenomena as sunspots, chromospheric eruptions, solar coronal disturbances, and ionospheric and geomagnetic disturbances are indices but not direct measurements. Although expressed in terms of tidal forces, Pettersson's celestial equation is also an expression of probable systematic changes in the amount of selective solar radiation reaching the earth's upper atmosphere. Proponents of the solar emission theory point to the proposed indirect or direct correlations between sunspots, terrestrial magnetism, atmospheric pressure patterns, and atmospheric tides in support of this theory. It is reasonable to suggest that the changes in tidal intensity, solar insolation, and other geophysical mechanisms implicit in the Pettersson curve may also affect pressures and circulation pattern changes in the lower atmosphere through an unknown combination of thermal, kinetic, and electrochemical reactions in the upper atmosphere. Whatever the physical causative agency or agencies involved, the fact that there appears to be a direct correlation between the climatic and tidal-intensity curves over the past tens of thousands of years (not just for the last 5000 years as assumed by Brooks) indicates that these may be directly related through atmospheric rather than through oceanographic mechanisms.

The Milankovitch-Pettersson Climatic Theory

Of all the solar climatic theories under consideration, the Milankovitch-Pettersson theory based on the geometric variations of solar insolation and related geophysical phenomena is the only one whose theoretical elements are sufficiently known to permit direct testing by comparison with the dated chronology of past climatic changes. Demonstration of a parallelism between these theoretical curves, dated by celestial mechanics, and the paleoclimatic record should provide pragmatic proof of a causal relationship whose ultimate verification, however, must be established through a quantitative meteorological and geophysical demonstration of the physical mechanisms linking the upper atmosphere phenomena with the machinery of shifting lower atmospheric circulation patterns.

Comparison of the paleoclimatic data with the solar insolation curve of Milankovitch and the tide-generating force curve of Pettersson as summarized in FIGURE 6 reveals no fundamental incompatibility between these astronomic periodicities and the climatic record. Instead, interrelations are suggested that apparently integrate the astronomic and climatic sequences into a unified astro-climatic system explaining not only the Pleistocene glacial-interglacial cycle of stage rank but also subordinate cycles of substage and lesser rank.

The inferred harmonic or near-harmonic series of astroclimatic cycles as expressed in years as simple ratios of the average glacial substage cycle is as follows:

40,800 ± (obliquity cycle = glacial stage cycle = 12 × substage cycle);
20,400 ± (precession cycle = 6 × substage cycle);
ca. 3400 (substage cycle);
1700 ± (Pettersson cycle = $\frac{1}{2}$ of substage cycle);
1133 ± (stadial cycle = $\frac{1}{3}$ of substage cycle);
567 ± ($\frac{1}{6}$ of substage cycle); and
283 ± ($\frac{1}{12}$ of substage cycle).

If this inferred cyclical series is valid, there is a sound theoretical basis for considering that other cycles of smaller, intermediate, and larger wave lengths may be found in the climatic record along with possible counterpart astronomic cycles in the dynamic system of planetary movements.

The difficulty and significance of establishing a quantitative meteorological and geophysical proof of the Milankovitch-Pettersson climatic theory can hardly be overemphasized. Such a proof essentially requires a theoretical break-through in establishing the significant interrelations between upper atmosphere phenomena, at present but vaguely understood, and the lower atmospheric circulation patterns. Its great significance is that it would provide a theoretical basis not only for a construction of a sound geoclimatic classification of the Quaternary period, but also for valid predictions of both small and large climatic trends into the future.

Summary

Current climatic theories may be directly tested by paleoclimatic evidence relating to regional aspects of past climatic zonation and to the recorded pattern of past climatic changes. Compilation and regional synthesis of field observations on surficial deposits of Alaska made largely by U.S. Geological Survey personnel over the past decade have greatly increased our understanding of Pleistocene climatic environments and glacial history of Alaska. Mapping and radiocarbon-dating of glacial deposits in the Cook Inlet region have permitted the detailed reconstruction of an independently dated glacial chronology. This paper summarizes the Alaskan data in relation to independently dated paleoclimatic evidence from other regions, and assesses their bearing on current climatic theories.

It is concluded:

(1) That the pertinent Alaskan and collated paleoclimatic evidence is clearly incompatible, both regionally and chronologically, with the geologic model required by the Ewing and Donn climatic theory.

(2) That this evidence suggests a harmonic or near-harmonic system of climatic cycles with the following wave lengths expressed in years as simple ratios of the glacial substage cycle; 283 ± ($\frac{1}{12}$); 567 ± ($\frac{1}{6}$); 1133 ± ($\frac{1}{3}$); 1700 ± ($\frac{1}{2}$); ca. 3400 (substage cycle), 20,400 ± (6×); and 40,800 ± (12×). The 40,800 year glacial stage cycle is broadly in phase with the obliquity cycle (average 40,800); the 20,400 year pluvial (?) cycle with the precession cycle (average ca. 21,000 years); and the 283,567,1133 and 3400 year cycles are

harmonically in phase with the Pettersson tide-generating force cycle (average 1700 years). Thus there is a strong presumption of genetic relations between these astroclimatic oscillations theoretically derived from celestial mechanics and the inductively derived and independently dated paleoclimatic sequence. Furthermore, in keeping with astroclimatic theory, the obliquity cycle seemingly dominated the climatic sequences (glacial and associated evidence) in upper latitudes and the precession cycle the climatic sequences (varve and pluvial (?) evidence) in lower middle latitudes.

(3) The pertinent Alaskan and collated paleoclimatic evidence thus favors a theory of geometric variations of solar insolation and associated geophysical phenomena over the "sunspot theory" of intrinsic solar activity as the primary cause of past climatic oscillations. The possibility is suggested that some or all of the short-term sunspot and collated terrestrial cycles may prove to be but subcyclical manifestations of the same geometric harmonic system inferred above.

References

Andersen, S., H. DeVries & W. H. Zagwijn. 1960. Climatic change and radiocarbon dating in the Weichselian glacial of Denmark and the Netherlands. Geol. Mijnbouw. **39**(2): 38–41.

Bennema, J. 1953. Holocene movements of land and sea level in the coastal area of the Netherlands. : 254–262. *In* Symposium: Quaternary changes in level, especially in the Netherlands. Geol. Mijnbouw. **16**(6).

Bradley, W. 1929. The varves and climate of the Green River epoch: U. S. Geol. Survey Prof. Paper, Washington. **158E:** 87–110.

Brandtner, F. 1950. Über die relative Chronologie des jüngeren Pleistozäns Niederösterreichs. Archaeol. Austriaca. **5:** 1950.

Brandtner, F. 1954. Jungpleistozäner Löss und fossile Boden in Niederösterreich. Eiszeitalter u, Gegenwart. **4/5**.

Bretz, J. H. 1959. The double Calumet stage of Lake Chicago. J. Geol. **67**(6): 675–684.

Broecker, W. S. 1957. Evidence for a major climatic change close to 11,000 years B.P. Geol. Soc. Am. Bull. **68**(12): pt. 2, 1703.

Broecker, W. S., M. Ewing & B. C. Heezen. 1960. Evidence for an abrupt change in climate close to 11,000 years ago. Am. J. Sci. **258**(6): 429–443.

Broecker, W. S. & J. L. Kulp. 1957. Lamont natural radiocarbon measurements IV. Science. **126:** 1324–1334.

Broecker, W. S., J. L. Kulp & C. S. Tucek. 1956. Lamont radiocarbon measurements III. Science. **124:** 154–165.

Broecker, W. S. & P. C. Orr. 1956. Late Wisconsin history of Lake Lahontan. Geol. Soc. Am. Bull. **67**(12): pt. 2, p. 1675.

Broecker, W. S. & P. C. Orr. 1958. The radiocarbon chronologies of Lake Lahontan and Lake Bonneville. Geol. Soc. Am. Bull. **69:** 1009–1032.

Broecker, W. S., K. K. Turekian & B. C. Heezen. 1958. The relation of deep-sea sedimentation rates to variations in climate. Am. J. Sci. **256:** 503–517.

Brooks, C. E. P. 1949. Climate Through the Ages. McGraw-Hill. New York, N. Y.

Craig, B. G. & J. G. Fyles. 1960. Pleistocene geology of Arctic Canada. Geol. Survey of Canada Paper. **60**(10): 1–21.

Deevey, E. S., L. J. Gialenski & V. Hoffren. 1959. Yale natural radiocarbon measurements IV; Radiocarbon supplement. Am. J. Sci. **1:** 144–172.

deGeer, E. H. 1957. Old and new datings of Swedish ice lakes and the thermals of Bölling and Alleröd. Geol. Fören. Förhandl. Bd. **79**(1): 93–100.

deGeer, G. 1940. Geochronologia suecica principles. K. Svensk, Vet. Akad. Handl., Stockholm, 350 p., 90 pl.

deVries, H., H. T. Barendsen & H. T. Waterbolk. 1958. Groningen radiocarbon dates II. Science. **127:** 129–137.

Douglass, A. E. 1936. Climatic cycles and tree growth. Carnegie Inst. Publ. no. 289, Washington.

Dreimanis, A. 1958. Beginning of the Nipissing phase of Lake Huron. J. Geol. **66**(5): 591–594.

DREIMANIS, A. 1960. Pre-classical Wisconsin in the eastern portion of the Great Lakes region, North America. Report of the XXI session Norden, Proceedings of section 4, p. 108–119.

EARDLEY, A. J., V. GVOSDETSKY & R. E. MARSELL. 1957. Hydrology of Lake Bonneville and sediments and soils of its basin. Geol. Soc. Am. Bull. **68**(9): 1141–1202.

EMILIANI, C. 1955. Pleistocene temperatures. J. Geol. **63**(6): 538–578.

EMILIANI, C. 1958. Paleotemperature analysis of core 280 and Pleistocene correlations. J. Geol. **66**(3): 264–275.

EMILIANI, C. & J. GEISS. 1957. On glaciations and their causes. Geol. Rundschau. **46**(2): 576–601.

ERICSON, D. B. & G. WOLLIN. 1956. Micropaleontological and isotopic determinations of Pleistocene climate. Micropaleontology. **2**(3): 257–269.

ERICSON, D. B. & G. WOLLIN. 1959. Micropaleontology and lithology of Arctic sediment cores. Geophysical Research Papers. **1**(63): 51–58.

ERICSON, D. B. 1961. Atlantic deep-sea sediment cores. Geol. Soc. Am. **72**(2): 193–286.

EWING, M. & W. L. DONN. 1956. A theory of the ice ages. Science. **123**: 1061.

EWING, M. & W. L. DONN. 1958*a*. A theory of the ice ages II. Science. **127**(3307): 1159–1162.

EWING, M. & W. L. DONN. 1958*b*. *In* Betty Friedman. The Coming Ice Age. Harpers magazine. Sept. : 39–45.

FAIRBRIDGE, R. W. 1958. Dating the latest movements of the Quaternary sea level. Trans. N. Y. Acad. Sci. Ser. 2. **20**: 471–482.

FAIRBRIDGE, R. W. 1960. Long range correlation of solar variation and sea level cycles. Technical Report, Office of Naval Research Contract, NONR 266 (69).

FAIRBRIDGE, R. W. 1961. Convergence of evidence on climatic change and ice ages. N. Y. Acad. Sci. **95**(1): 177–188.

FETH, J. H. & M. RUBIN. 1957. Radiocarbon dating of the wave-formed tufas from the Bonneville basin. Geol. Soc. Am. Bull. **68**(12): pt. 2: 1827.

FISK, H. N. & E. MCFARLAN, JR. 1955. Late Quaternary deltaic deposits of the Mississippi River. : 279–302. *In* A. Poldervaarte, Ed. Crust of the earth. Geol. Soc. Am. Special Paper. **62**: 762 p.

FLINT, R. F. 1947. Glacial Geology and the Pleistocene Epoch. 6 pl. Wiley. New York, N. Y.

FLINT, R. F. 1953. Evidence from glacial geology as to climate variations: p. 165–178. *In* Climatic Change. Harvard Univ. Press. Cambridge, Mass.

FLINT, R. F. 1956. New radiocarbon dates and late Pleistocene stratigraphy. Am. J. Sci. **254**: 265–287.

FLINT, R. F. 1957. Glacial and Pleistocene Geology. Wiley. New York, N. Y.

FLINT, R. F. & F. BRANDTNER. 1961. Climatic changes since the last interglacial. Am. J. Sci. **259**: 321–328.

FLINT, R. F. & W. A. GALE. 1958. Stratigraphy and radiocarbon dates at Searles Lake, California. Am. J. Sci. **256**(10): 689–713.

FLINT, R. F. & M. RUBIN. 1955. Radiocarbon dates of pre-Mankato events in eastern and central North America. Science. **121**: 649–658.

FREY, D. G. 1953. Regional aspects of the late-glacial and post-glacial pollen succession of southern North Carolina. Ecological monographs. **23**: 289–313.

FREY, D. G. 1955. A time revision of the Pleistocene pollen chronology of southeastern North Carolina. Ecology. **36**(4): 762–763.

FROMM, E. 1960. An interglacial peat at Ale near Luleå, northern Sweden. Sveriges Geol. Undersökning. **54**(5): 3–14.

FRYE, J. C. & H. B. WILLMAN. 1960. Classification of Wisconsin stage in the Lake Michigan glacial lake. Ill. State Geol. Survey Circ. **285**: 1–15.

GODWIN, H., D. WALKER & E. H. WILLIS. 1957. Radiocarbon dating and post-glacial vegetational history. Scale by Moss. Proc. Roy. Soc. London. **B147**: 352–366.

GODWIN, H. & E. H. WILLIS. 1959. Radiocarbon dating of the late-glacial period in Britain. Proc. Roy. Soc. **150**: 199–215.

GOLDTHWAIT, R. P. 1958. Wisconsin age forests in western Ohio. Ohio J. Sci. **58**(4): 209–219.

GRANLUND, E. 1932. De Svenska högmossarnas geologi. Sver. Geol. unders. Arsbok. **26**(1): 1–193.

GROSS, H. 1958. Die bisherigen Ergebnisse von C 14 Messungen und Paläontologischen Untersuchungen für die Gliederung und Chronologie des Jungpleistozans in Mitteleuropa und den Nachbargebieten. Eiszeitalter und Gegenwart. **9**: 155–187.

HANSEN, S. 1940. Varvighedni danskeiogrskaanske senglaciale A flejringer. Danmarke Geol. Undersøgelse. **11**(63): 413–477.

Hough, J. L. 1953. Pleistocene climatic record in a Pacific Ocean core sample. J. Geol. **61**(3): 252–262.
Hough, J. L. 1958. Geology of the Great Lakes. Univ. Ill. Press. Urbana, Ill.
Iversen, J. 1954. The late-glacial flora of Denmark and its relation to climate and soil. : 87–119. *In* Studies in vegetational history in honour of Knud Jessen. Geol. Survey of Denmark, **11,** Ser. 80.
Järnefors, B. & E. Fromm. 1960. Chronology of the ice recession through middle Sweden. Proc. XXI Int. Geol. Congr. **IV:** 93–97.
Jennings, J. D. 1957. Danger Cave: Mem. 14, Soc. Am. Archaeology. *In* Am. Antiquity. **23,** pt. 2. (also in Utah Univ. Dept. Anthropology. Anthropol. Papers no. 27, 328 p.).
Karlstrom, T. N. V. 1955. Late Pleistocene and Recent glacial chronology of south-central Alaska (abs.). Geol. Soc. Am. Bull. **66:** 1581–1582.
Karlstrom, T. N. V. 1956. The problem of the Cochrane in late Pleistocene chronology. U. S. Geol. Survey Bull. **1021J:** 303–331.
Karlstrom, T. N. V. 1957*a*. Tentative correlation of Alaskan glacial sequences. Science. **125**(3237): 73–74.
Karlstrom, T. N. V. 1957*b*. Alaskan evidence in support of a post-Illinoian pre-Wisconsin glaciation (abs.). Geol. Soc. Am. Bull. **68:** 1906.
Karlstrom, T. N. V. 1959. Reassessment of radiocarbon dating and correlations of standard late Pleistocene chronologies (abs.). Geol. Soc. Am. Bull. **70**(12): pt. 2: 1627.
Karlstrom, T. N. V. 1960*a*. The Cook Inlet, Alaska, glacial record and Quaternary classification. U.S. Geol. Survey Prof. Paper **400B:** 330–332.
Karlstrom, T. N. V. 1960*b*. Surficial deposits of Alaska. U.S. Geol. Survey Prof. Paper. **400B:** 333–335.
Karlstrom, T. N. V. *et al.* 1959. Surficial deposits of Alaska. U.S. Geol. Survey open-file report. Washington, D.C.
Kempton, J. P. & R. P. Goldthwait. 1959. Glacial outwash terraces of the Hocking and Sciota River valleys, Ohio. Ohio J. Sci. **59:** pt. 3, 135–151.
Klíma, B. 1958. Příspěvek ke stratigrafii nejmladšího sprašového pokryvu (Abs: A contribution to the stratigraphy of the youngest loess cover). Anthropozoikum. **7**(1957): 111–143. 10 pl. Praha, 1958.
Lawrence, D. B. 1950. Glacial fluctuation for six centuries in southeastern Alaska and its relation to solar activity. Geol. Rev. **40:** 191–223.
Lee, H. A. 1960. Late glacial and postglacial Hudson Bay sea episode. Science. **131**(3413): 1609–1611.
Leighton, M. M. 1958. Important elements in the classification of the Wisconsin glacial stage. J. Geol. **66**(3): 288–309.
Leighton, M. M. 1960. The classification of the Wisconsin glacial stage of northcentral United States. J. Geol. **68**(5): 529–552.
Libby, W. F. 1955. Radiocarbon Dating. Univ. Chicago Press. Chicago, Ill.
Livingston, D. A., M. Ewing & W. L. Donn. 1959. Theory of ice ages. Science. **129**(3347): 464–465.
Lundqvist, G. 1956. C-14 metoden Kvartärgeologien. Ymer, Stockholm, Sweden. : 231–236.
Magnusson, N. H., E. Granlund & G. Lundqvist. 1947. Sveriges geologi Svenska Bokforlaget. Stockholm, Norstedts., 424 p.
Milankovitch, M. 1941. Kanon der Erdbestrahlung und seine Andwendung auf das Eiszeitenproblem. Academie Royal Serbe, Editions Speciales, Tome CXXXIII, Section des Sciences Mathematiques et Naturelles, Tome 33, Belgrade.
Morrison, R. B. 1952. Stratigraphy of Lake Lahontan and associated Quaternary deposits in the Carson Desert area near Fallon, Nevada. Geol. Soc. Am. Bull. **63**(12): pt. 2: 1367.
Nilsson, E. 1960. The recession of the land ice in Sweden during the Alleröd and the Younger Dryas ages. Proc. XXI Intern. Geol. Congr. Part IV. : 98–107.
Olson, E. A. & W. S. Broecker. 1959. Lamont natural radiocarbon measurements V. Am. J. Sci. Radiocarbon Supplement. **1:** 506.
Pelišek, J. 1958. Kvatérni sedimenty paleolitické stanice V Ostravě-Petrkovicich. Anthropozoikum. **8:** 61–71, pl. I and II (Quaternary sediments of Ostrava-Petrovice; with a short German summary).
Pettersson, O. 1914. Climatic variations in historic and prehistoric time. Svenska Hydrogr. Biol. Komm. Skriften., **5.**
Piggot, C. S. & W. D. Urry. 1942. Time relations in ocean sediments. Geol. Soc. Am. Bull. **53:** 1187–1210.
Post, L. von. 1946. The prospect for pollen analysis in the study of the earth's climatic history. The New Phytologist. **45**(2): 193–217. (Vega Lecture, 1944. Prologue from p. 212.)

PREST, V. K. 1957. Pleistocene geology and surficial deposits, p. 443–495. *In* Stockwell, Geology and Economic Minerals of Canada. Geol. Survey of Canada Economic Ser. No. 1.

RAY, L. L. 1960. Significance of loess deposits along the Ohio River valley. U.S. Geol. Survey Prof. Paper. **400B:** 211.

RICHMOND, G. M., R. B. MORRISON & H. J. BISSELL. 1952. Correlation of the late Quaternary deposits of the LaSal Mountains, Utah, and of the Lake Bonneville and Lahontan by means of interstadial soils. Geol. Soc. Am. Bull. **63**(12): pt. 2, 1369.

ROOSMA, A. 1958. A climatic record from Searles Lake, California. Science. **128**(3326): 716.

RUBIN, M. & C. ALEXANDER. 1958. U.S. Geological Survey radiocarbon dates IV. Science. **127**(3313): 1476–1487.

RUBIN, M. & C. ALEXANDER. 1960. U.S. Geological Survey radiocarbon dates V. Radiocarbon Supplement, **2.** Am. J. Sci. : 129–185.

RUHE, R. V., M. RUBIN & W. H. SCHOLTES. 1958. Late Pleistocene radiocarbon chronology in Iowa. Am. J. Sci. **255:** 671–689.

SCHOVE, D. J. 1955. Sunspot cycle, 649 B.C. to A.D. 2000. J. Geophys. Research. **60**(2): 127–140.

SCHULMAN, E. 1956. Dendroclimatic Changes in Semiarid America. Univ. Ariz. Press. Tucson, Ariz.

TERASMAE, J. & O. L. HUGHES. 1960. Glacial retreat in the North Bay area, Ontario. Science. **131**(3411): 1444–1445.

THORARINSSON, S. 1940. Present glacial shrinkage and eustatic changes of sea level. Geogr. Ann. Stockholm. **22:** 131–159.

TROWBRIDGE, A. C. 1954. Mississippi River and Gulf coast terraces and sediments as related to Pleistocene history—a problem. Geol. Soc. Am. Bull. **65**(8): 793–812.

URRY, W. D. 1942. The radio-elements in non-equilibrium systems. Am. J. Sci. **240:** 426–436.

WILLETT, H. C. 1953. Atmospheric circulation. : 52–71. *In* Climatic Changes. Harvard Univ. Press. Cambridge, Mass.

WILLIS, E. H., H. TAUBER & K. O. MÜNNICH. 1960. Variations in the atmospheric radiocarbon concentration over the past 1300 years. Am. J. Sci. Supplement. **2:** 1–4.

WISEMAN, J. D. H. 1958. La Topographie et la géologie des profondeurs oceaniques. Colloques Internationaux Centre National de la Recherche Scientifique. **83:** 193–208.

WISEMAN, J. D. H. 1959. The relation between paleotemperatures and carbonate in an equatorial Atlantic pilot core. J. Geol. **67**(6): 685–690.

WOERKOM, A. J. J. VON. 1953. The astronomic theory of climatic change. *In* Climatic Change. Harvard Univ. Press. Cambridge, Mass.

ZEIST, W. VAN. 1958. Some radiocarbon dates from the raised bog near Emmen (Netherlands). Palaeohistoria. **iv:** 113–118.

ZEUNER, F. E. 1950. Dating the Past. 475 p., 24 pl. Methuen. London, England.

ZUMBERGE, J. H. 1960. Correlation of Wisconsin drifts in Illinois, Indiana, Michigan and Ohio. Geol. Soc. Am. Bull. **71**(8): 1177–1188.

ZUMBERGE, J. H. & J. E. POTZGER. 1956. Late Wisconsin chronology of the Lake Michigan basin correlated with pollen studies. Geol. Soc. Am. Bull. **67**(3): 271–288.

RESPONSE OF ENCLOSED LAKES TO CURRENT GLACIOPLUVIAL CLIMATIC CONDITIONS IN MIDDLE LATITUDE WESTERN NORTH AMERICA*

Donald B. Lawrence and Elizabeth G. Lawrence
University of Minnesota, Minneapolis, Minn.

Recent studies of glacier variation in western United States have shown that renewed growth began about 1942 after several decades of accelerated recession. The new tendency was first detected at higher levels of Nisqually Glacier on Mount Rainier, Wash., in 1944 by Johnson (1954). A spectacular advance of the Coleman Glacier on Mount Baker apparently began about 1949 as noted by Bengtson (1956). Spasmodic growth of Lyell Glacier in the central Sierra Nevada of California occurred as early as 1932, according to Harrison (1956). By the middle of the past decade the trend had become so general that Hubley (1956) was able to record enlargement in 50 glaciers of the Olympic Mountains and Cascade Range of Washington between 1953 and 1955, associated with lowered temperature and increased precipitation. Subsequent observations of Mark F. Meier indicating growth of still other glaciers of the Cascade Range, especially the Forsyth on Mount St. Helens with an advance of 2100 feet in the interval between 1938 and 1958, have been reported in a newspaper account by Richards (1959). Growth of glaciers on Mount Hood, Ore., between 1947 and 1956 has been demonstrated by Handewith (1959 and by unpublished information, Mazama Mountaineering Club of Portland, Ore.).

The trend of recent glacier growth has also been observed northward in British Columbia and Alaska, as reported by Austin S. Post (personal communication), who has indicated that "In 1960 probably well over one hundred small, steep, gradient glaciers in the coast range between Vancouver, B.C. and Juneau, Alaska either were advancing or casting ice down cliffs where older ice was no longer present." Miller (1960) has similarly pointed out that the zone of maximum snowfall on the Juneau Icefield in southeastern Alaska has taken a strong downward shift during the last years of this decade.

Concurrently with the growth of these western glaciers, a general rise has been noted in the levels of certain enclosed lakes. These lakes have no surface outflow; hence, like the glaciers, they enlarge or diminish with changes in precipitation and evaporation. In Oregon, water levels had risen by 1958 to nearly all-time record stages on Abert, Crater, Davis, East, Goose, and Silver Lakes (Fairbanks, 1954; and unpublished information, United States Geological Survey and National Park Service, Washington, D.C.). In central Washington, the marked rise of Soap Lake has been ascribed to climatic causes (Mundorff and Bodhaine, 1954). The studies of enclosed lakes (reported earlier by Harding, 1949) and the wish to extend our own field observations in British Columbia and eastern Washington that began in 1938 led to further study afield in the summer of 1960.

* The field work involved in preparing this study was made possible by aid granted by the Graduate School of the University of Minnesota.

Twenty lakes were examined (some of those in FIGURE 1) between northern California and central British Columbia. Although organization of the resulting data has scarcely begun, it seems worthwhile to present this preliminary report of the methods and also tentative conclusions as to the synchronism of glacial and pluvial climates and the hydrologic effects of small climatic changes.

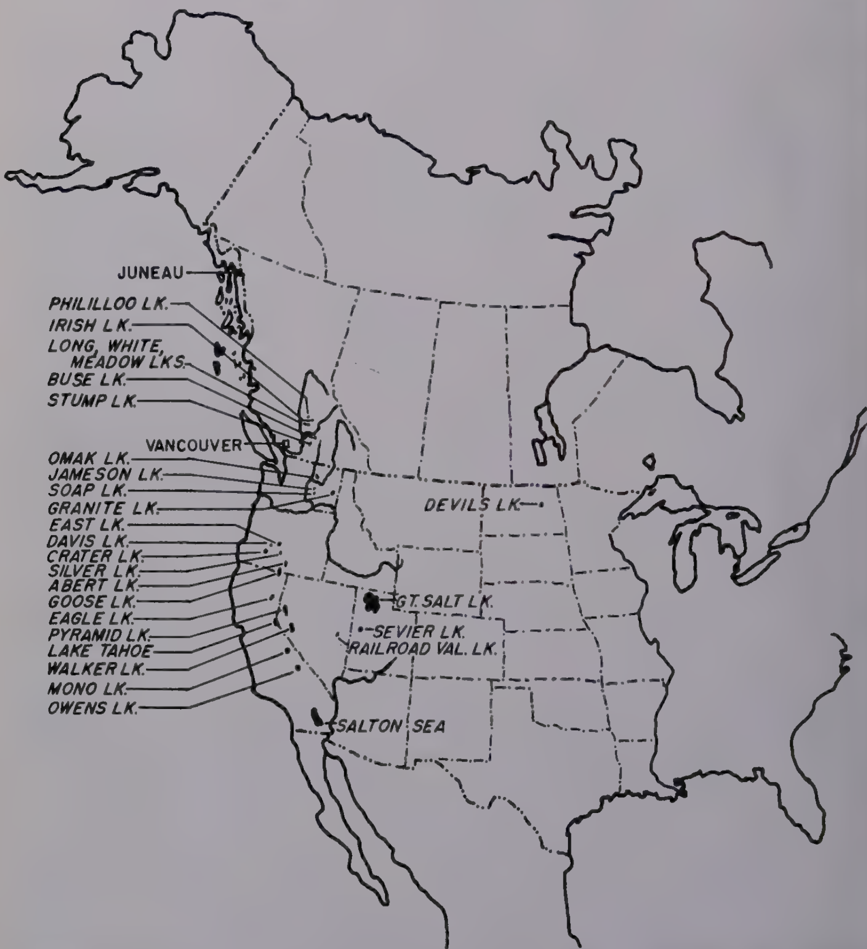


FIGURE 1. Some closed basin lakes of western North America whose water-level histories are being studied.

In our studies we attempted to learn as much as possible about the history of changes in level of water surface in enclosed basins. We measured the heights above current water surfaces of the bases of drowned trees, of old living trees, of high water lines on rocks, and of shorelines formed by wave erosion and deposition during past high-water stages. We sounded depth of water over the root systems of trees drowned at various times and, at some lakes, of living trees rooted well below current water stages. To ascertain

time relationships we sampled for age certain critical individual trees and shrubs, both drowned and living, with Swedish increment borers and saws. We obtained chronologic data at some places by study of living trees that had been tilted in the 1880s by wave erosion at the highest lake levels of the past century, and at other places by cross-dating between drowned and old living trees. Wherever possible we interviewed long-time local residents to assemble human historical data on the lake level changes for comparison with our data based on botanical evidence. We obtained all available official gauge readings. Regional climatic trends as inferred from tree ring studies of Douglass (1919 to 1936), Keen (1937), Antevs (1938), and Schulman (1956) are of special value for the prehistoric period. In several areas we reoccupied photographic stations established in 1938 and 1941 in the early stages of these studies and obtained new series of pictures. By comparing these sets of photos taken 22 years and 19 years apart, water level changes can be seen and evaluated.

Through the study area from northern California in the south to central British Columbia in the north, preliminary analysis of the data reveals the following consistent trends: high levels in the 1870s and 1880s and again in the early 1900s, then extreme desiccation until the middle and late 1930s, when some basins were completely dry. Water levels began to rise rather steadily in the late 1930s or early 1940s, and the rise became accelerated in the 1950s. Rises of as much as 16 feet have occurred in the past two decades in many basins, with broad extension of lake surface areas and the drowning of such vegetation as trees, shrubs, and lichens, many of them 30 to 50 years old. Goose Lake, for example, at an elevation of 4700 feet on the California-Oregon boundary, increased from a dry flat in 1930 to about 160 sq. mi. by 1958; as early as 1936 open water had begun to persist through the summer on the basin floor. This increase in level occurred in spite of gradually increasing use of tributary streams for irrigation.

To the southeast in Nevada, the levels of Walker Lake at a 4000-foot elevation and Pyramid Lake at a 3800-foot elevation have fallen sharply through the period of regular record between 1927 and 1959, 65 feet at Walker and 45 feet at Pyramid. This progressive lowering may have resulted from increased use of tributary streams and ground water for irrigation, or from greater importance of evaporation as contrasted with precipitation in that part of the region, or both. However, even here there are suggestions that the general trend noted farther north has had its effects. For Pyramid Lake, the highest observed level of the calendar year exceeded that of the preceding year 7 times in the period 1938 to 1958, and at Walker Lake the peak in 1938 exceeded that of 1937. Great Salt Lake in Utah had its recent crest in 1952.

Several of the lakes studied are not affected by storage, diversion, or the works of man. Examples are Crater, Davis, and East Lakes in Oregon, Omak and Granite Lakes in Washington, and Stump and Buse Lakes in British Columbia. These all have followed a definite pattern: high stages in the first decade or two of the 20th century, then a steady drop to about 1935 to 1941, followed by a rapid rise to very high stages in 1958 or 1960. Most of the lakes affected by diversions for irrigation show a parallel response. At the more northerly lakes, highest levels tended to occur in 1960, so that in that

summer it was usual in Washington and British Columbia to find flooded forests, pastures, fences, roads, and buildings.

It was interesting to find that the residents of each area considered their experience with the rising water to be unique, and various hypotheses had been developed locally to explain it. None included the possibility of a general climatic change. The rises were attributed to supposed causes such as local cloud seeding, silting of submerged outflow seepage channels, inward seepage over long distances, and even uphill in some cases from irrigation reservoirs, conversion of native range to agricultural crops, and logging of adjacent forests. Some of these influences may well have contributed to the rise of waters; but the broad geographical extent of the trend is surely too great for the general validity of any one of them.

Geographic distribution of the trend is not yet precisely defined, but it is known to extend eastward at middle latitudes to include Devils Lake, N.D. (Swenson and Colby, 1955, and recent unpublished data), and the basins of the upper Mississippi (Isaak *et al.*, 1959) and the Red and St. Lawrence Rivers (R. S. Goodridge, 1960, personal communication).

One of our observations was of special interest to us because we had been watching for evidence of a high-lake stage that might have been contemporary with the most extensive recent glacier advance of the late 17th and early 18th centuries. We had previously found such advance on Mount Hood and farther northward in Alaska (Lawrence, 1950, 1958), and others have reported it in Norway and elsewhere in the northern hemisphere. That brief glaciation seems to have been induced by reduced solar activity associated with the great 70-year sunspot dearth period of 1645 to 1715 (FIGURE 2) which was independently discovered by Douglass (1919, p. 102) and by Maunder (1921–22, pp. 140 to 145). This same period of sunspot dearth and a number of other earlier ones as far back as the first century A.D. have also been reported by Kanda (1933). At Davis Lake, at a 4390-foot elevation in Oregon, we found an old sand bar lying across the mouth of a cove (FIGURE 3) with its level crest about 2.2 feet *above* the high-water mark of 1957, the highest of recent years. A prominent boulder rampart formed by ice action when the lake stood at the high sand-bar level was also found. We could be sure that the sand bar had been formed at a high-water stage that had occurred more than 235 years ago because on its crest was rooted the stump of a recently living ponderosa pine that had been cut within two or three years. This pine had 235 growth rings on the stump surface 1 foot above the root crown. Appearances suggested that this had been one of the first trees to become established at about 1720 A.D. on the then newly formed sand bar. A high water line that stood at the same level as the top of the old sand bar and may also have been formed at that same time was still discernible (FIGURE 4*a*) on rocks of the lava flow that had dammed the valley to form the lake. Its position could be detected because lichen colonies of a different kind and color grew above and below the old water line (FIGURE 4*b*).

We have seen that recent physical changes both in glaciers and enclosed basin lakes have been simultaneous and parallel. For many years scientists have been wondering whether the great fossil lakes (FIGURE 5) in regions that

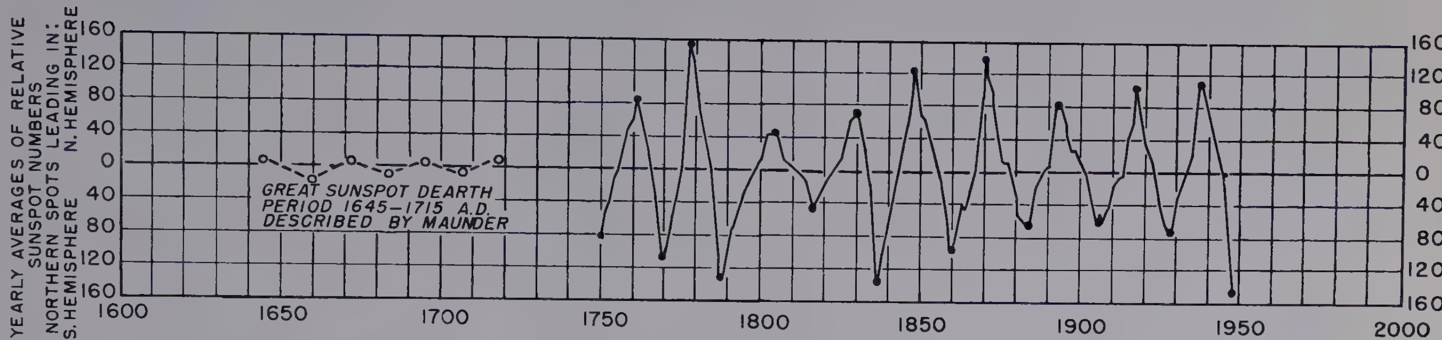


FIGURE 2. Graph of the sunspot cycle with alternating peaks plotted above and below a zero line as was first done by Anderson (1939) to show successive reversals in orientation of magnetic polarity of leading and following members of sunspot pairs in succeeding 11-year periods. Observed values of annual average relative sunspot numbers are plotted as the solid line connecting solid circles. The broken line at left connecting open circles shows Maunder's "great sunspot dearth period" 1645 to 1714 A.D., based on reduced occurrences of sunspots and auroras reported in the astronomical literature. Many glaciers grew to greatest size of recent centuries during or immediately following the 70-year period of reduced solar activity. Some evidence is accumulating now which suggests that high stages in enclosed basin lakes occurred at the same time. Reproduced (modified) by permission of the *American Scientist* (Lawrence, 1958).

are now desert were formed at the same time that glaciers grew to such extraordinary size as to cover large parts of northern North America. It had been assumed that these occurrences were synchronous, and there is general evidence for that assumption. For example, radiocarbon dates recently available suggest that the glacier and lake growth were synchronous; but the events occurred so long ago that the dating must be general, so that it is difficult to be sure of synchronism. Furthermore it has not been possible to find out what causal mechanisms operated so long ago. The recent simultaneity of growth of lakes and adjacent glaciers seems to help satisfy the need pointed

FIGURE 3. Cove at northeast edge of Davis Lake, Ore. Ponderosa pines about 240 years old are rooted on crest of sand bar formed at a high lake stage of approximately 4,395.4 feet elevation, a little before 1720 A.D. Lava flow that blocked the valley and formed the lake is plainly visible. Lawrence photo of August 14, 1960. The photos of FIGURE 4 were taken just beyond right edge of this picture.

out by Hubbs and Miller (1948, p. 119) for more accurate dating of hydrographic history and more definite correlation with montane glaciation.

It would seem that these events of recent decades have made it possible to apply modern meteorological concepts and analyses to some problems regarding the mechanics of climatic change. Through the study of concrete and rather well-documented examples such as this one of the latest 20 years in comparison with records for the previous 2 decades, it is reasonable to expect that we may find out why certain sets of years are so different from others. When we do, we may well be on the path toward new understanding of causes of glacial and interglacial periods, as Faegri (1950) suggested. Faegri expressed the belief that the changes that brought on the ice ages were no more extreme than the differences we ourselves have experienced between extreme years, but that the changes merely lasted longer then.

FIGURE 4. (a) Surveyor's rod against lava flow at northeast edge of Davis Lake, Ore., with base set at high water line of June 1957. Note white stain on rocks below base of rod. (b) Detail of light-colored crustose lichens above old high-water line at about 2.2 feet on the rod, and black foliose lichens below. This line stands at the same elevation as the crest of the sand bar seen in the foreground of FIGURE 3, and probably was formed at the same time, about 1720 A.D. Lawrence photos of August 14, 1960.

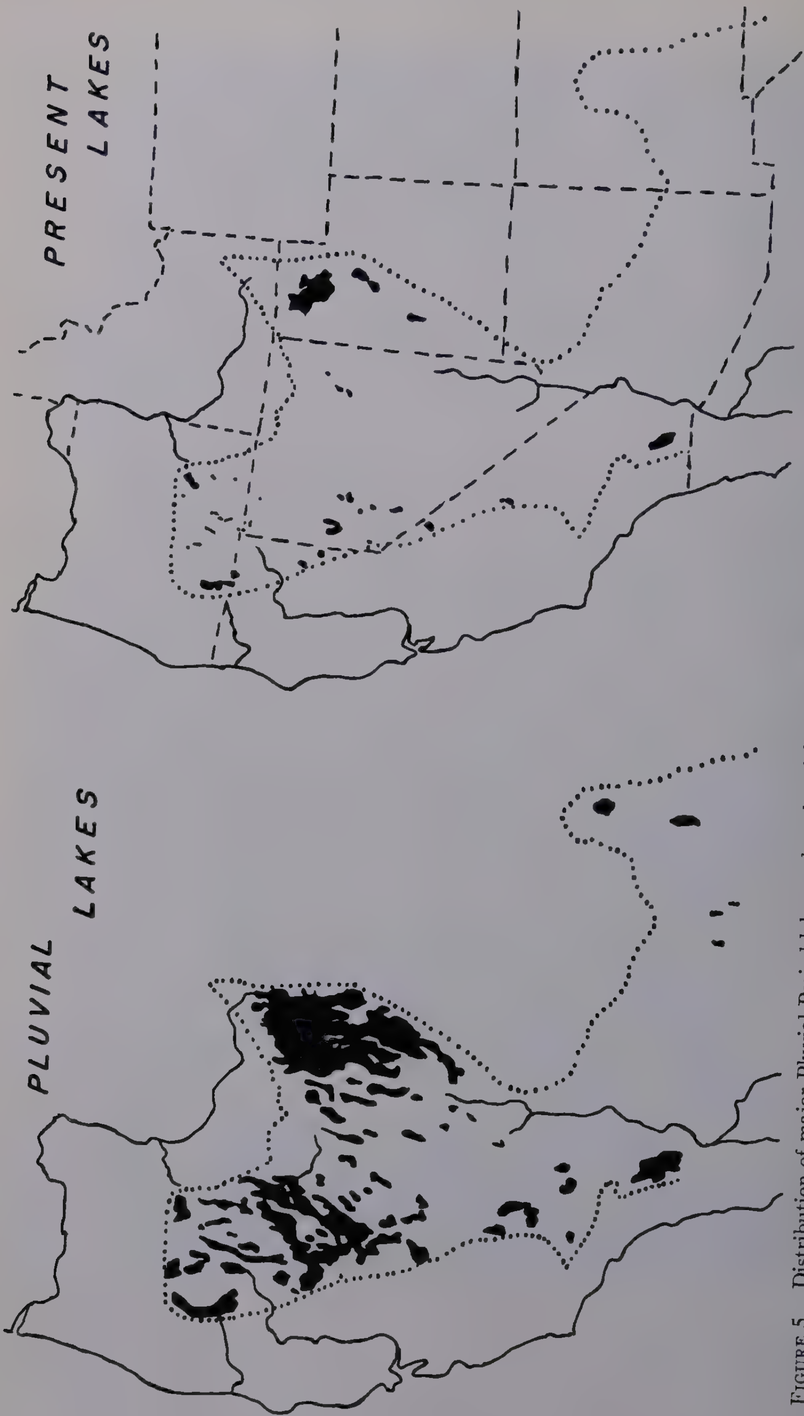


FIGURE 5. Distribution of major Pluvial Period lakes and modern lakes in western North America. Reproduced (modified) by permission of the *Bulletin of the Geological Society of America* (Meinzer, 1922).

Summary

A survey of the history of fluctuations of 20 closed basin lakes from northern California to central British Columbia showed strong growth since 1940 with crests ranging from 1952 to the summer of the survey, 1960. The more recent crests, chiefly in 1957, 1958, and 1960, tended to occur in the more northerly parts of the area. The simultaneous growth of glaciers and lakes in the latest 20 years from shrunken conditions of the previous 20 suggests that a study of the weather differences in these periods may reveal some information as to the origin of glaciopluvial climates of the past.

Acknowledgments

We express our gratitude to K. N. Phillips, of the United States Geological Survey, and to Bruce Black, Crater Lake National Park, for aid in the field in Oregon. For making available unpublished data on lake levels we thank Professor Emeritus S. T. Harding of the University of California, Berkeley, Calif., and the following members of the United States Geological Survey: F. M. Veatch of Tacoma, Wash., L. J. Snell of Carson City, Nev., M. T. Wilson of Salt Lake City, Utah, H. M. Erskine of Bismarck, N.D., L. R. Sawyer of St. Paul, Minn. and, especially, K. N. Phillips of Portland, Ore., who also assisted significantly in the preparation of this paper. We are grateful to Roy Strickland, State of Washington Department of Game, at Omak, Wash., and Carey K. Eastwood, Soil Conservation Service, at Waterville, Wash., for unpublished data on Jameson and Omak Lakes.

References

ANDERSON, C. N. 1939. A representation of the sunspot cycle. Bell-Syst. Tech. J. **18:** 292–299.

ANTEVS, E. 1938. Rainfall and tree growth in the Great Basin. Carnegie Inst. Washington Pub. No. **469**, and Am. Geog. Soc. Spec. Pub. No. **21**. 97 p. + 2 plates.

BENGTSON, K. B. 1956. Activity of the Coleman Glacier, Mount Baker, Washington, U. S. A., 1949–55. J. Glaciol. **2**(20): 708–713.

DOUGLASS, A. E. 1919–1936. Climatic cycles and tree growth. 3 vols. Carnegie Inst. Pub. No. **289**.

FAEGRI, K. 1950. On the value of paleoclimatological evidence. Centenary Proc. Roy. Meteorol. Soc. **1950:** 188–195.

FAIRBANKS, C. W. 1954. Crater Lake waters. U. S. D. I. National Park Service. Crater Lake Nature Notes **20:** 30–34.

HANDEWITH, H., JR. 1959. Recent glacier variations on Mount Hood. Mazama. **41**(13): 23–28. See also errata sheet accompanying Mazama **42** (13), 1960. (Portland, Oregon).

HARDING, S. T. 1949. Lakes. Chapt. **6:** 220–243. *In* Hydrology (Physics of the Earth—IX). O. E. Meinzer, Ed. Dover. New York, N. Y.

HARRISON, A. E. 1956. Glacial activity in the western United States. J. Glaciol. **2**(19): 666–668.

HARRISON, A. E. 1956. Fluctuations of the Nisqually Glacier, Mt. Rainier, Washington, since 1750. J. Glaciol. **2**(19): 675–683.

HUBLEY, R. C. 1956. Glaciers of the Washington Cascade and Olympic Mountains; their present activity and its relation to local climatic conditions. J. Glaciol. **2**(19): 669–674.

HUBBS, C. L. & R. R. MILLER. 1948. The zoological evidence. *In* The Great Basin with emphasis on Glacial and Postglacial Times. Bull. Univ. Utah. **38**(20): 17–166.

ISAAK, D., W. H. MARSHALL & M. F. BUELL. 1959. A record of reverse plant succession in a tamarack bog. Ecology. **40:** 317–320.

JOHNSON, A. 1954. Observations on the nisqually glacier and other glaciers in the Northwestern United States. Union Geodesique et Geophysique Internationale. Assoc. Internat. d'Hydrol. Sci. Assemblee gen. de Rome, 1954: Tom. 4, Comptes-rendus et Rapports de la Commiss. de Neiges et des Glaces. : 511–516.

KANDA, S. 1933. Ancient records of sunspots and auroras in the Far East and the variation of the period of solar activity. Proc. Imperial Acad. (Japan). **9:** 293–296

KEEN, F. P. 1937. Climatic cycles in eastern Oregon as indicated by tree rings. Monthly Weather Rev. **65:** 175–188.
LAWRENCE, D. B. 1950. Glacier fluctuation for six centuries in Southeastern Alaska and its relation to solar activity. Geog. Rev. **40:** 191–223.
LAWRENCE, D. B. 1958. Glaciers and vegetation in Southeastern Alaska. Am. Scientist. **46:** 89–122.
MAUNDER, E. M. 1921–22. The prolonged sunspot minimum, 1645–1715. J. Brit. Astron. Assoc. **32:** 140–145.
MEINZER, O. E. 1922. Map of Pleistocene lakes of the Basin-and-Range Province and its significance. Geol. Soc. Am. Bull. **33:** 541–552.
MILLER, M. M. 1960. Status memo on the Juneau Icefield Research Program, summer season, 1960. Maynard M. Miller, Department of Geology, Michigan State University, East Lansing, Mich. 2 p. (ditto).
MUNDORFF, M. J. & G. L. BODHAINE. 1954. Investigation of the rise in level of Soap Lake at Soap Lake, Washington. Open-file Report. Water Resources Div. U. S. Geological Survey, Tacoma District. 44 p. plus two plates.
RICHARDS, L. 1959. Advancing Northwest glaciers store vast water supply visioned as future source of hydroelectric energy. The Oregonian. Sunday, Nov. 8, 1959. 6M page 28.
SCHULMAN, E. 1956. Dendroclimatic changes in semiarid America. Univ. Ariz. Press. Tucson, Ariz.
SWENSON, H. A. & B. R. COLBY. 1955. Chemical quality of surface waters, Devils Lake Basin (North Dakota). U. S. Geological Survey Water Supply Paper 1295. U. S. Govt. Printing Office. Washington, D.C.

CONNECTIONS BETWEEN THE SECULAR VARIATIONS OF THE EARTH'S MAGNETIC FIELD AND OTHER PHENOMENA

G. Barta

Roland Eötvös Geophysical Institute, Budapest, Hungary

Data from the magnetic observatories of the Northern Hemisphere have revealed the existence of a periodic fluctuation of about 40 to 50 years in the secular variation of the earth's magnetic field. It is interesting to note that the rotational speed of the earth's axis shows a similar oscillation. This may be an indication of the existence of a mass displacement of the same period in the interior of the earth, that is, in the core.

This mass movement may cause a change of the gravitational field also; therefore it may possess a secular variation which could be found in the data of the past 100 years as a period of 40 to 50 years. For lack of sufficiently accurate gravity measurements we may use the oscillations of the ocean level when investigating the gravitational variation, the surface of the quiet ocean being an equipotential surface of the gravitational field. Unfortunately, the ocean level is also influenced by meteorological factors, oceanic currents, and crustal movements; moreover, there are very few tide-recording stations possessing a sufficiently long series of data. In spite of this, we can detect a period of about 50 years in some series, as in the data for Aberdeen-Dunbar, Scotland; Sydney, Australia; La Goulette, Tunis; Bombay, India; and Honolulu Hawaii; and these series show a distinctly parallel, respectively antiparallel march, although these stations lie extremely far one from another. Thus we may conclude that the variation is a result of some far-reaching mass movement affecting the whole of the earth.

The comparison of the secular variations of the various observatories presents some difficulties. In order to eliminate the demonstrational difficulties, orthogonal projections were applied to the variational vectors, the projecting directions radiating from certain points. The investigations revealed the existence of a center of symmetry somewhere in Pakistan. This indicates the flow of a circular current below Pakistan to a depth of about 3000 km. As is known, the magnetic center of the earth lies eccentrically 340 km. in the direction of the Marshall Islands, and this center has been moving in the last century in WNW direction. We can conclude from the eddy formed around Pakistan that the inner core of the earth is moving toward Pakistan.

The identification of the magnetic center with the innermost core of the earth leads of course to other conclusions; a mass-eccentricity of such an extent might influence also the shape of the earth. Indeed, the major axis and oblateness of the equatorial ellipse of the tri-axial earth agree—within the limits of measuring accuracy—with the data of the ellipse that can be calculated from the direction and extent of the magnetic eccentricity.

The connections of the phenomena as indicated here may provide a possibility to make observations, by which we can obtain a verification of the hypothesis of an internal mass movement. For this purpose we should arrange relative g measurements on a closed network of stations with nearly

equal gravity values lying around the equator, and we should repeat the observations in a time interval of 5 to 10 years. If the innermost core of the earth is moving with the speed of the magnetic center, then after 10 years the occurrence of a double wave of 0.6 to 0.8 mgal amplitude in the series of the relative g values would be observed. The modern measuring accuracy provides the possibility of exact determination of even much smaller gravity values, thus making the project very promising.

Of course, a movement of the innermost core of the earth of such an extent can be brought about only by energies of adequate size. Let us investigate these energies. During the oscillation, manifesting itself in the rotational speed of the earth in a period of 50 years, the earth is either in advance, or in retardation, in relation to the average uniform movement, and the maximum accumulated value of this advance is $+1$ sec. at the end of the first quarter of the period, the corresponding accumulated value of the retardation being -1 sec. at the end of the third quarter of the oscillation. Thus we may express the amount of the whole rotation performed in the part of the period up to a given point marked t in the following formula:

$$\alpha = 2\pi t + \frac{2\pi}{86{,}400} \sin \frac{2\pi}{18{,}250} t$$

where α is the angle of rotation in radians; t is the time in days. From this we can get a formula for the rotational speed by differentiating against t.

It is easy to calculate the energy-fluctuation corresponding to this variation in speed; as a result of the calculation, we can state that in the phases with maximum rate of change in speed, the earth is gaining an amount of energy of 6.10^{24} ergs; respectively it loses such an amount daily. The order of magnitude of this immense energy is the same as that released in the course of earthquakes in one year. The energy of the sun radiated to the earth is even greater by several orders of magnitude; nevertheless it cannot influence directly the rotational speed. The total amount of wind energy of our atmosphere is $2{,}400 \cdot 10^{24}$ ergs, that is, 400 times as great as the energy appearing in the oscillations of the rotational speed; but the bulk of these movements is of an unsystematically varying direction. The speed oscillation could be brought about only by air masses moving around the Equator in an east-west direction; the energy transfer would be accomplished by the friction only. However, an oscillation of energy of such an extent ought to be observed in the amount of the wind speed near the ground.

Cosmical energies are much greater than meteorological ones; therefore, they may be used better in clarifying these phenomena.

The revolving energy of the moon is $E_M = 3.6 \cdot 10^{35}$ ergs.

The revolving energy of the earth is $E_E = 2.7 \cdot 10^{40}$ ergs.

The problem in this connection is: how could the energy be transferred from the heavenly bodies to the earth? If the earth is of a centrisymmetrical structure, then it is well balanced in the solar system; the eccentrical inner core, however, is under the influence of external forces, depending on the eccentricity. Let us assume that the inner core with its density of 17 is floating inside the outer core of a density of 11; then—taking into account the rota-

tion of the earth—we can calculate that the value of the acceleration caused by the moon to the earth, owing to the eccentricity of the core, amounts to $a_M = 2.0 \cdot 10^{-6}$ cgs.

The corresponding values for the sun and Jupiter are:

$$a_S = 0.9 \cdot 10^{-6},$$

$$a_J = 0.9 \cdot 10^{-8} \text{ cgs.}$$

The effect increases the eccentricity in all cases. As a result of this effect, the distance of the core from the axis of rotation is growing, its momentum of inertia increases and its rotational speed relative to the crust diminishes; therefore it suffers a displacement against W. Owing to this movement toward W the development of the above-mentioned symmetry feature of the magnetic secular variation—with its center about Pakistan—is taking place.

The earth, the moon, and Jupiter are moving along elliptical paths. The effects of the forces are 4 to 5 per cent greater at the ends of the minor axes than at the ends of the major ones; therefore the astronomical periods must present themselves in the secular variation also. The effects of the sun and moon amplify one another if the direction against the perigee of the moon's path is opposing the direction of the perihelion of the earth's path. Thus these effects had been added in 1881, 1916, and 1951, in good agreement with the oscillation experienced in the earth's rotational speed and in the magnetic secular variation.

We have seen that astronomical energies have their part, in all probability, in determining the magnetic secular variation; on the other hand, one may ask whether the magnetic secular variation could exert any influence on our climate.

The general circulation of our atmosphere is governed by the sun's radiation of heat. The energy forwarded by the corpuscular radiation of the sun is of a much smaller dimension, and its distribution in space and time is also of a wholly different character. The wave radiation of the sun arrives at our earth continuously on the day side, chiefly around the equator and in the temperate latitudes, according to the conditions set by the geometrical shape of the earth. The energy of the corpuscular radiation, on the other hand, reaches us with a widely varying intensity in the polar regions and on the night side. It is this difference in the distribution in space and time that may produce the possibility that these energies of relatively small dimensions may trigger high energy processes and so they may have some role in regulating the development of our weather mechanism.

That is where some interrelation to the magnetic secular variation could be expected, as the magnetic field of the earth may have an influence upon the quantity and place of arrival of the energies carried by the sun's corpuscular radiation. Thus the development of the general circulation of our atmosphere may be influenced, too, by magnetic variations, although the intervention of some mediatory agents seems to be necessary.

VARIATIONS IN THE GENERAL ATMOSPHERIC AND HYDROSPHERIC CIRCULATION OF PERIODS OF A FEW YEARS DURATION AFFECTED BY VARIATIONS OF SOLAR ACTIVITY

H. P. Berlage

Royal Netherlands Meteorological Institute, De Bilt, the Netherlands

The essence of the present conference is the show of evidence that relations exist between the fluctuations of world weather and the fluctuations of the intensity and the composition of solar radiation. My contribution is on the relatively short-period oscillation in world weather that was named the "Southern Oscillation" (S.O.) by Sir Gilbert Walker. The S.O. is an oscillation of the intensity of the general atmospheric and hydrospheric circulation through the Pacific and Indian oceans, the water hemisphere of our globe. The two main centers of operation, the scales of the air-pressure balance, are the subtropic high-pressure area around Easter Island (27° 10′ S, 109° 26′ W) and the equatorial low-pressure area of Indonesia.

We may designate Djakarta, formerly Batavia (6° 11′ S, 106° 50′ E), as the station representative of this Malay Low. FIGURE 1 is a graph of the six-monthly running means of the air-pressure anomalies at Djakarta, 1866 to 1960. The air-pressure oscillations show a 2 to 3 year rhythm. The physical reality of this periodicity is assured by TABLE 1. It contains the frequency distribution of intervals in halves of years between successive minima. This distribution is such that randomness seems to be excluded.

The most promising interpretation of the S.O. is the following.[1] If air-pressure is abnormally low in the Malay area it is abnormally high in the Easter Island area. An abnormally high pressure gradient between the subtropical high and the equatorial low strengthens the general atmospheric circulation. Consequently the hydrospheric circulation is accelerated. The Peru Current is stronger than normal. The South Equatorial Current is accelerated. The water masses arriving in the Malay area are colder than normal. Consequently sea surface temperatures in Indonesia are decreased. Low sea-surface temperatures cause a rise of air pressure in the Malay area. If air pressure is abnormally high in the Malay area it is abnormally low in the Easter Island area. An abnormally low pressure gradient between the subtropical high and the equatorial low slackens the general atmospheric circulation. Consequently the hydrospheric circulation is decelerated. The Peru Current is weaker than normal and hence warmer than normal. The South Equatorial Current is retarded. The water masses arriving in the Malay area are warmer than normal. Consequently sea-surface temperatures in Indonesia are increased. High sea-surface temperatures cause a fall of air pressure in the Malay area. Thus the cycle is closed.

If, as in the present case, in one and the same area (1) the temperature anomaly follows the air-pressure anomaly with a time lag averaging 7 months, and (2) the time derivative of the air-pressure anomaly is inversely proportional to the temperature anomaly, an oscillation of air pressure and temperature must occur of, roughly, a 4 × 7-month or between 2- and 3-year period.[1]

What remains an open question is the upper air motion between the two centers. That upper air plays a major role in the S.O. is proved by the fact that air-pressure fluctuations of this periodicity are as high at mountain stations, for example, Mount Pangerango in Java, as they are at sea level.[2] Vertical wind profiles observed at Christmas Island in the Pacific Ocean suggest

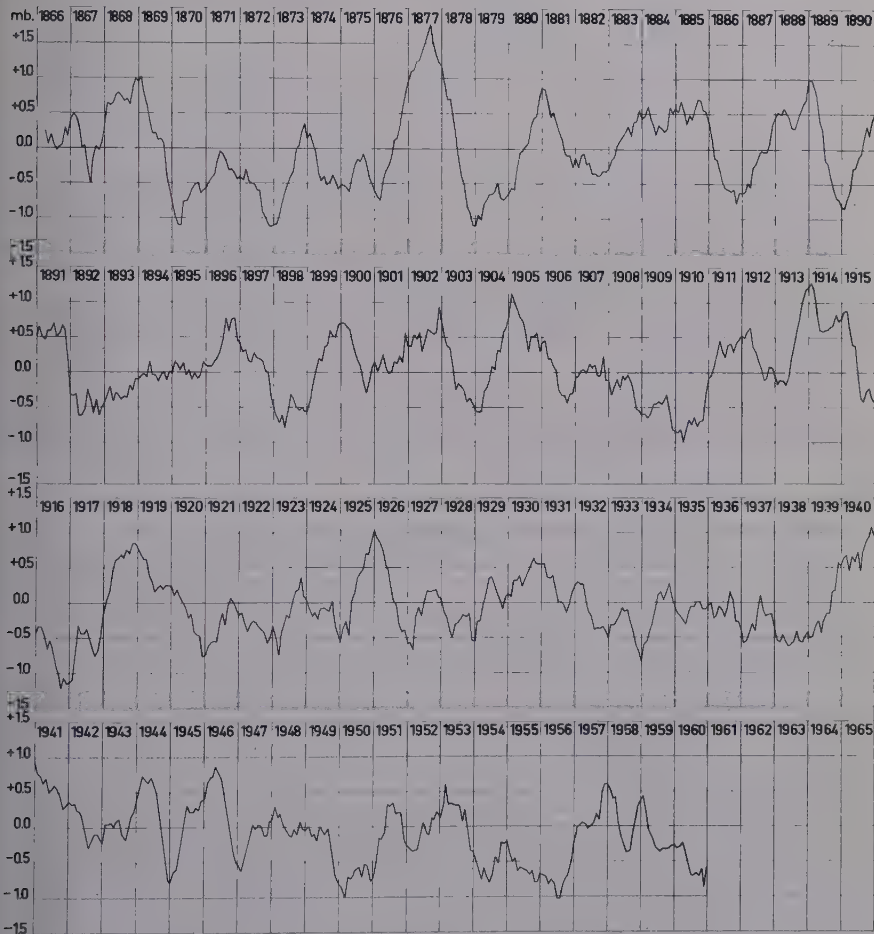


FIGURE 1. Djakarta air-pressure fluctuations; running six-monthly departures of mean 1866 to 1940.

TABLE 1

PERCENTAGES OF FREQUENCY OF INTERVALS BETWEEN MINIMA OF DJAKARTA SEMIANNUAL AIR-PRESSURE DEVIATIONS IN HALVES OF YEARS

	2	3	4	5	6	7	8
Empirical	18	21	27	23	9	0	2
Expected	40	33	17	7	2	0.6	0.13

that upper air directions are actually reversed from east to west in the course of one cycle of the S.O.[3,4]

What we have explained thus far is that sea-level conditions generate the S.O. It is a typical relaxation effect of terrestrial origin. Even continuous inflow of solar radiation of the same intensity and composition would lead to a S.O. In order to detect solar variations in the S.O. we must look for the variations occurring in its period and amplitude.

FIGURE 2 contains besides the Djakarta air-pressure curve and the air-pressure curve of the island Juan Fernandez (33° 37′ S, 78° 52′ W). This station was established in 1911. It is not more than 500 miles away from the continent, and in its latitude traveling cyclones are already making so much "noise" as to hide S.O. components of the air-pressure fluctuations.[1] It is definitely less representative for the subtropical high than Easter Island, but we are obliged to use its data through the years in which Easter Island observations are missing. Moreover these data are at least better than those of continentally situated Santiago (33° 27′ S, 70° 42′ W).

We learn from FIGURE 2 that there are definite phase lags between the Juan Fernandez and the Djakarta air-pressure waves. One aspect of the S.O. is its standing wave character, another aspect is its progressive wave character. It is apparently not always the case that a Juan Fernandez extreme precedes the opposite Djakarta extreme. This is, however, mostly true, and the priority of Juan Fernandez is perhaps most prominent where very large waves occur that very often also show large amplitudes.

These cases and their outstanding importance explain the curious fact that correlation between Djakarta and Juan Fernandez anomalies is positive and greatest at an average interval of 2 to 2½ years; curious in so far as such a lag would suggest[1] an average S.O. period of 4 to 5 years. This suggestion, however, should not deceive us. The significance of these very long and high waves will be discussed presently.

FIGURE 3 shows us first how remarkably intimate the relation is between sea-surface temperature at Puerto Chicama (7° 47′ S, 79° 28′ W) and air pressure at almost antipodal Djakarta. The normal course of both elements is evidently simply parallel. The literally global extension of the S.O. is herewith clearly demonstrated. A pressure maximum at Djakarta is contemporaneous with a pressure minimum at Easter Island. This means a minimum of the air-pressure gradient Easter Island-Djakarta, a minimum driving force and a deceleration of the hydrospheric circulation. The Peru Current is slow; hence abnormally warm along the Peru coast. A pressure minimum at Djakarta is contemporaneous with a pressure maximum at Easter Island. This means a maximum of the air-pressure gradient Easter Island-Djakarta, a maximum driving force and an acceleration of the general atmospheric circulation. This causes an acceleration of the general hydrospheric circulation. The Peru Current is quick and, accordingly, abnormally cold along the Peru coast.

The phase lag between cause and effect may explain the usual slow rise and quick fall of air pressure at Djakarta that is the saw-tooth character of the S.O. Moreover, a high pressure gradient Easter Island-Djakarta shortens

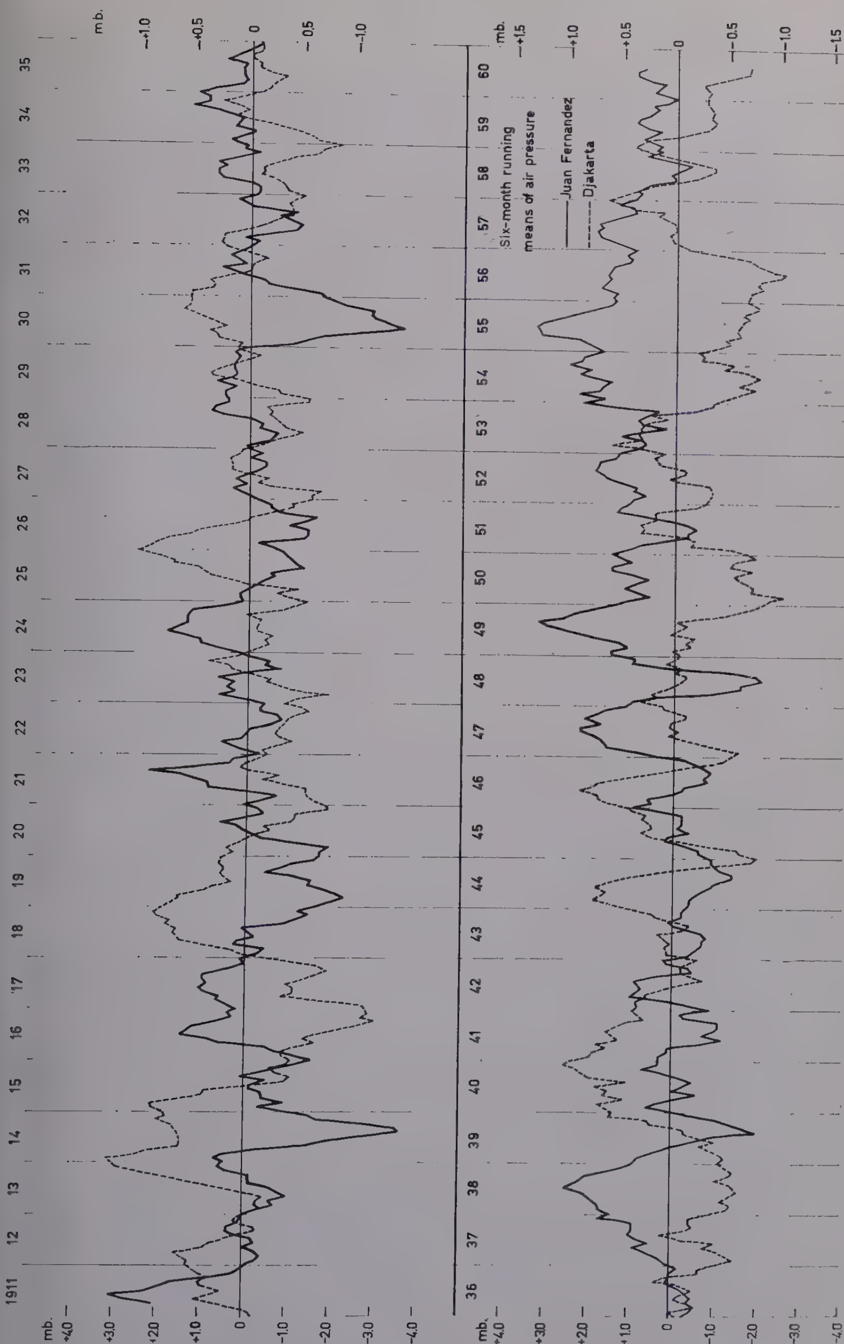


FIGURE 2. Djakarta (*right scale*) and Juan Fernandez (*left scale*) running six-monthly anomalies of air pressure.

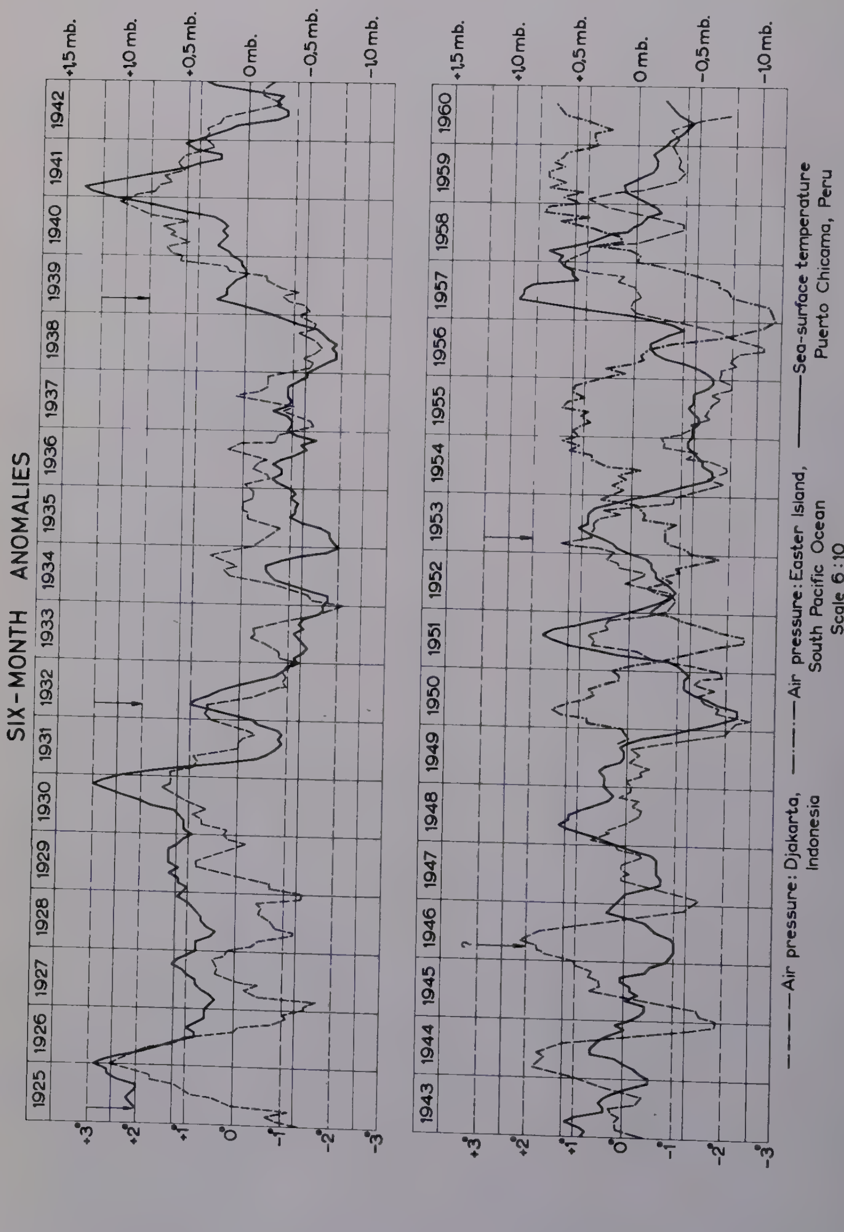


FIGURE 3. Djakarta and Easter Island running six-monthly anomalies of air pressure, and Puerto Chicama running six-monthly anomalies of sea-surface temperature.

the period of the S.O.; a low pressure gradient Easter Island-Djakarta lengthens the period of the S.O.

FIGURE 3 presents also evidence of the operation of a more-or-less independent 7-year cycle in sea surface temperatures. It has long been known from the remarkable El Niño appearances.

The El Niño current is the outbreak from north to south along the Peru coast of warm equatorial water in the southern summer. We note 1925, a year of real El Niño catastrophe, and further sharp sea-surface temperature peaks in 1932 and 1939. A 1946 sea-surface temperature maximum remained absent, very characteristically. Hydrospheric circulation and, probably, also atmospheric circulation must have been anomalous in 1946. Air-pressure observations at Easter Island showed a maximum that year coincident with the Djakarta maximum. The S.O. behaved abnormally, at least in the eastern South Pacific.

In 1953 the El Niño sea-surface temperature peak occurred regularly. If I am right, air pressure should be quickly rising now at Djakarta, quickly falling now at Easter Island and El Niño threatening, that is, in 1961 instead of in 1960. In fact, this slip is in accordance with the following rule, which seems to govern the remarkable appearances of El Niño at intervals that occur chiefly at 7 years, but which may be at 6 or 8 years. "A year in which the sunspot number decreases below 13 may become an El Niño year, after which a sequence of El Niño years at 7-year intervals is to follow. If one of these El Niño years is preceded or followed by a year in which the sunspot number decreases below 13 this latter year takes over the leading role from the former and becomes the first of a new sequence."[1]

This rule is verified since 1807, a year in which the sunspot number dropped below 13, without exception. However the actual phenomenon has sometimes failed. It is associated with high pressure in Djakarta with the same restrictions. For instance, 1946 was a Djakarta high-pressure year, whereas 1898 was decidedly not, although it is known as an El Niño year.

Closer inspection of FIGURE 1 suggests also solar influences on the S.O. proper. During sunspot maximum years 1870, 1871, 1872–1882, 1883, 1884–1892, 1893, 1894–1905, 1906, 1907–1916, 1917, 1918–1927, 1928, 1929–1937, 1938, 1939–1947, 1948, 1949–1957, 1958, and 1959 the Djakarta air-pressure curve seems relatively depressed; during sunspot minimum years 1866, 1867, 1868–1877, 1878, 1879–1888, 1889, 1890–1900, 1901, 1902–1911, 1912, 1913–1922, 1923, 1924–1932, 1933, 1934–1942, 1943, 1944-1952, 1953, and 1954 the Djakarta air-pressure curve seems relatively raised.

In fact this rule is true in only two thirds of the cases. There is no question of simple parallelism; 11-year or 22- to 23-year periodicities are poorly developed. However, solar variations are undoubtedly affecting the pace and the amplitude of the S.O. Trial and error induced me to stress the probability that the S.O. operates in the Indonesian sector according to the following 4 rules:

(1) The first high-pressure year after a sunspot minimum year occurs whenever the sunspot number exceeds 20. It follows 3 years after the previous high-pressure year if the sunspot number 20 has not yet been reached.

(2) The second high-pressure year follows 3, 4, 5, or 6 years later. It follows 3 years later if the sunspot number has not reached 81. It follows 4, 5, or 6 years later when the sunspot maximum has passed and the sunspot number has decreased below 81. If it follows 5 or 6 years later a secondary air-pressure maximum is developed.

(3) The third high-pressure year follows 2 or 3 years later: 2 years later when the sunspot number has decreased below 54; 3 years later when the sunspot number has not yet decreased below 54.

(4) The fourth high-pressure year follows 2 or 3 years later: 2 years later when the sunspot number has decreased below 13; 3 years later when the sunspot number has not yet decreased below 13.

These rules provide us with the average scheme of operation of the S.O., the mean annual sunspot number being the only variable. The relative independence of the 7-year El Niño cycle—perhaps simply 3 times the S.O. cycle—urges us to look for its intervention. An adequate rule of intervention in the Indonesian sector appears to be the following.

"A high-pressure year of the 7-year cycle, which is preceded and followed by a high-pressure year of the 2–3 year cycle remains a high-pressure year. If a high-pressure year of the longer cycle is preceded or followed by a high-pressure year of the shorter cycle, they are combined into one and it is the later year that becomes the high-pressure year."[1]

FIGURE 4 contains two graphs. The one represents the successive October–March and April–September average air-pressure values at Djakarta and the values that our rules provide. The persistence of the correlation between both curves in successive intervals is promising. It is shown by TABLE 2.

The total correlation through the whole series works out at +0.46. Since successively 4 tops are identified, the rules really amount to 4 independent ones. In this case +0.46 is the correlation coefficient that just reaches the level of statistical significance. This certainly is a disappointing statement. The great deceptions are in the 1860s and 1870s and in the 1930s and 1940s.

However it should be noted that the rules cited were derived primarily to account for the succession of abnormally dry and abnormally wet dry seasons, May to October, in Java. The rules do this with slightly greater reliability than when indicating the succession of high-pressure and low-pressure waves in Indonesia, which are for the greater part associated with relatively dry and relatively wet east monsoons in Java.

Moreover, from FIGURE 5, showing in addition to the Djakarta air-pressure fluctuations the annual anomaly of the air-pressure gradient Santiago-Djakarta, we get the strong impression that some of the apparent failures can be very well understood if the longer period fluctuations of the amplitude and period of the S.O. are again interpreted as relaxation oscillations. The gigantic wave-producing high pressure at Djakarta, 1940 to 1941, is preceded by 8 years in succession, during which the air-pressure gradient Santiago-Djakarta was below normal. The former station, being continental, has serious defects as representative of the subtropic high, but it is the only one available through all the years of Djakarta observations.

The general circulation through the Pacific and Indian oceans was accelerated during 8 years in succession: 1931 to 1938. As a matter of fact the S.O.

in the course of these years shows diminishing period and amplitude. Evidently a kind of threshold value is surpassed and the acceleration closed by an extremely high and long air-pressure and temperature wave. The next air-pressure peak arrives 3½ years after the first. Of course the sunspot maximum depression 1937–1938–1939 at Djakarta contributed to this effect.

An analogous case happened from 1870 onward. After 6 years of air pressure below normal at Djakarta and above normal at Santiago, the general circulation through the Pacific and Indian oceans having been speeded up beyond a certain critical limit, the enormous 1877 air-pressure wave occurs, the next peak arriving again 3½ years later. This time it is the sunspot maximum depression 1870–1871-1872 at Djakarta that has contributed a great deal to the total acceleration of the general circulation.

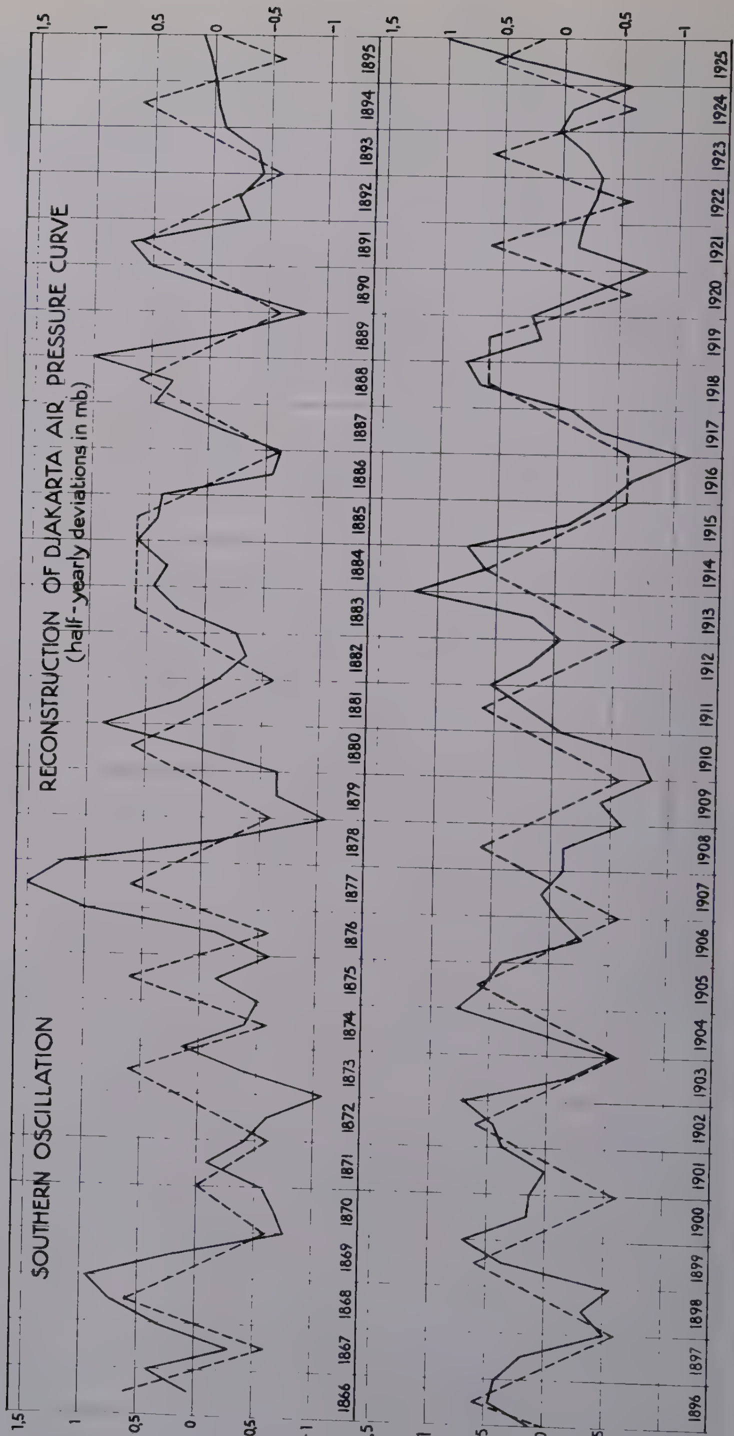
The third interesting case is the positive anomaly of the air-pressure gradient Santiago-Djakarta maintained through the years 1906 to 1913 and followed by the large air-pressure wave with peaks at Djakarta in 1914 and 1918. These occurred even at 4-year interval.

The latest case of this kind is the positive anomaly of the air-pressure gradient Santiago-Djakarta maintained through the years 1952 to 1956. The long wave following is the one showing a maximum at Djakarta 1957 to 1958 and a minimum 1959 to 1960. The next air-pressure maximum at Djakarta is evidently due in the course of 1961. This year almost certainly will present a sunspot number below 81; hence the coming air-pressure maximum at Djakarta will occur again 3 to 4 years after the previous one.

The reason why the 1957 to 1958 air-pressure peak in Djakarta rose only slightly above normal probably is that it coincides with the highest sunspot maximum ever recorded. Our rules would have required a secondary maximum only. That air pressure at Djakarta reached a level well above normal during these years is already remarkable, and a clear indication of the exceptional relaxation wave height.

My colleague H. J. de Boer is now studying a theory that might enable us to forecast, at given passages of the S.O. through its mean stage, the type of the forthcoming wave. Our first impression is that the cumulative anomaly of the air-pressure gradient Santiago-Djakarta (substituting the air-pressure gradient Easter Island-Djakarta) through a number of past years is the indicator for the new start. This might have enabled us to foresee the arrival of a 1939–1940–1941 S.O. wave at the beginning of 1939 or even earlier. Juan Fernandez air-pressure fluctuations, as shown by FIGURE 2, suggest that the relaxation came as early as 1937, the year in which extremely low temperatures were recorded in Lambayeque and Iquique, that is, in the eastern South Pacific. The reverse task is to forecast the introduction of such troubled periods as the one of 1931 to 1938. Perhaps in this case the warning signal has been given by the quite abnormally low temperatures recorded in 1930 in the western South Pacific.

The cases cited here seriously distort the provisional operation scheme of the S.O. following the rules based on sunspot numbers, in the 1930s and the 1940s. Even in 1940–1941–1942 two waves apparently united. Much is gained, however, if only we understand such transformations. Equally serious distortions occurred in the 1860s and 1870s previously mentioned, but again our knowl-


SOUTHERN OSCILLATION
RECONSTRUCTION OF DJAKARTA AIR PRESSURE CURVE
(half-yearly deviations in mb)

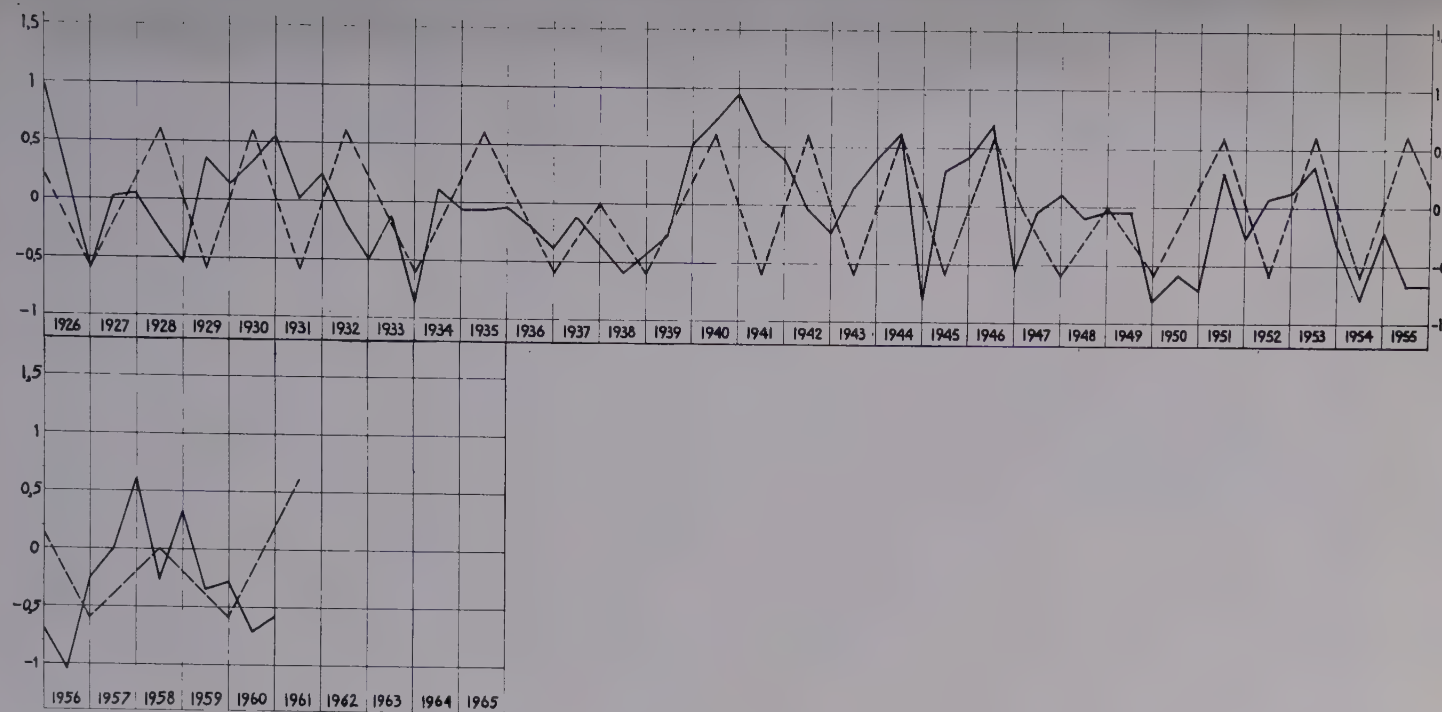


FIGURE 4. October to March and April to August average Djakarta air-pressure anomalies and their construction from annual sunspot numbers.

TABLE 2

COEFFICIENTS OF CORRELATION BETWEEN BOMBAY-DJAKARTA SEMIANNUAL AIR-PRESSURE ANOMALIES AND VALUES DERIVED FROM ANNUAL SUNSPOT NUMBERS

Bombay	1847–1856	+0.63
	1857–1866	+0.38
Djakarta	1866–1875	+0.25
	1876–1885	+0.58
	1886–1895	+0.57
	1896–1905	+0.51
	1906–1915	+0.42
	1916–1925	+0.51
	1926–1935	+0.11
	1936–1945	+0.25
	1946–1955	+0.35

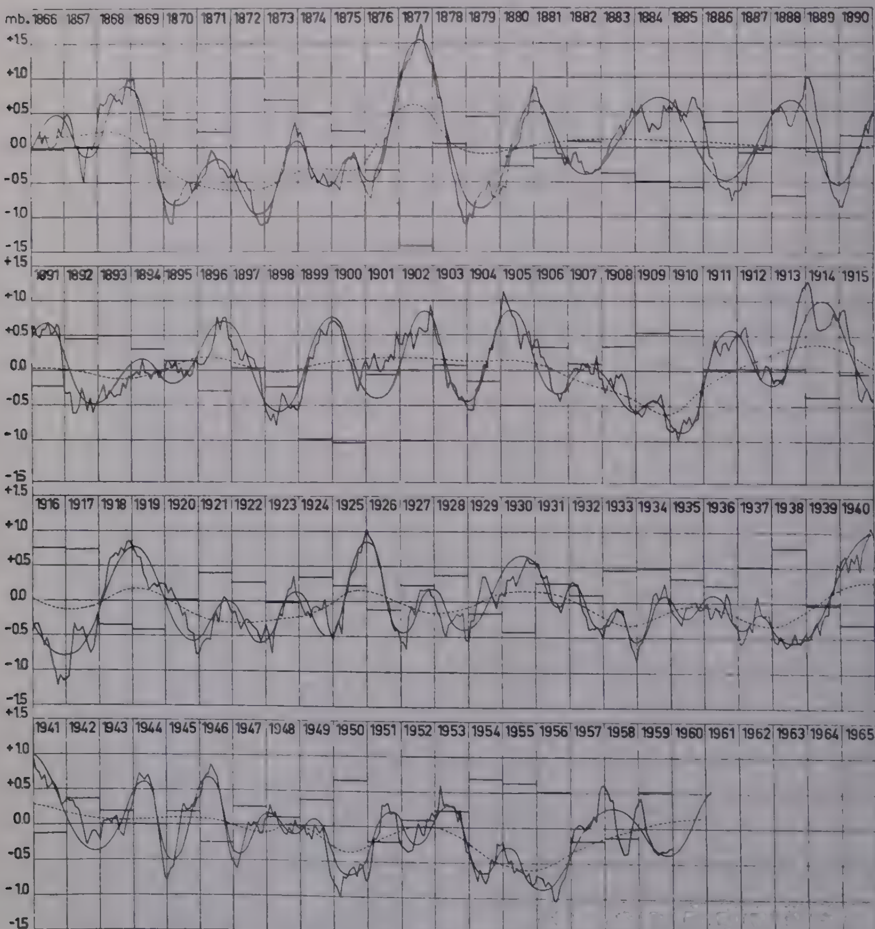


FIGURE 5. Running six-monthly departures from normal of Djakarta air pressure: dotted curve drawn through the middle heights of successive waves. Horizontal lines indicate the annual anomaly of Santiago-Djakarta air-pressure gradient.[5] Reproduced by permission of *Geofisica Pura e Applicata.*

FIGURE 6. Iso-correlates of twelve-monthly air-pressure anomalies with respect to Djakarta, January 1, 1931 to January 1, 1939.[5] Reproduced by permission of *Geofisica Pura e Applicata*.

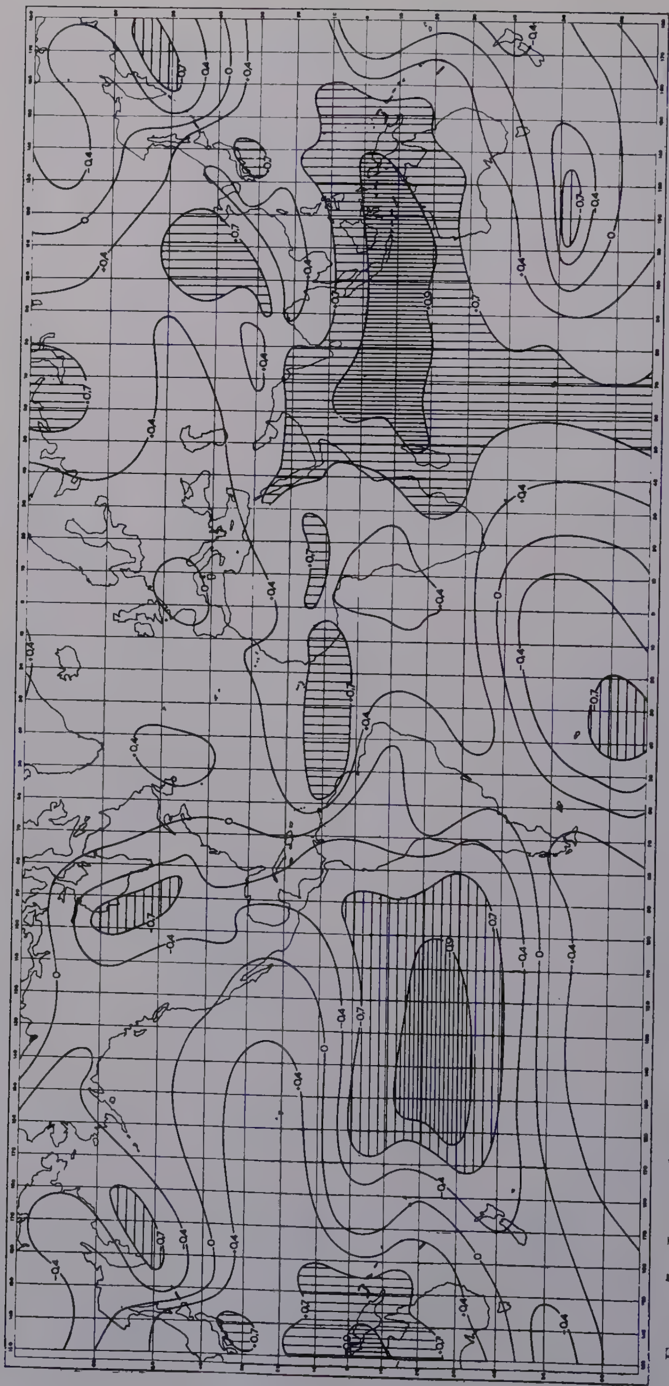


FIGURE 7. Iso-correlates of twelve-monthly air-pressure anomalies with respect to Djakarta, July 1, 1949 to July 1, 1957.[5] Reproduced by permission of *Geofisica Pura e Applicata*.

edge of their causes is increasing. Hence I am convinced that much can be done to ameliorate the correlation between the empirical and the constructed air-pressure curves during those anomalous intervals: the more so since a few secondary S.O. waves strongly suggest their dependence on the interception of solar radiation by world-wide volcanic dust veils, for example from Krakatoa (1883), Katmai (1912), and Bezumiany (1956).

The lines of this investigation converge toward the recognition of real solar terrestrial relationships. It is therefore also interesting to see what follows from a comparison of the world patterns of correlation of yearly pressure anomalies relative to Djakarta in 2 charts involving 2 characteristic series of years, 1931 to 1938 and 1949 to 1957 (FIGURES 6 and 7). When these charts are compared with 2 charts presenting these correlations during a much longer series of years, but separately for the seasons October to February and April to August,[5] we are able to show that the interval 1931 to 1938 has summer characteristics in both the Northern and Southern Hemispheres, whereas the interval 1949 to 1957 has winter characteristics. This would mean in a sense above-normal insolation of our globe in 1931 to 1938 and below normal insolation in 1949 to 1957. As an interesting instance of the characteristic difference between these 2 series of years I may mention the fact that all summers (June–July–August) from 1931 through 1939 were relatively dry in the Netherlands, whereas the majority of the summers from 1949 through 1957 in the Netherlands were wet to very wet. An acceleration of the S.O. such as occurred in 1931 to 1938 may go so far as to reduce its period from 30 months to perhaps 18 months.[5] In cases such as these the S.O. does not remain as dominant as usual. It is apparently reacted upon by the North Atlantic oscillation, whose proper period is 21 months on the average. During such intervals dominancy seems to be transferred from the S.O. to the N.A.O.; hence a definite feedback from the one oscillation to the other is demonstrated.

References

1. BERLAGE, H. P. 1957. Fluctuations of the general atmospheric circulation of more than one year, their nature and prognostic value. K.N.M.I. De Bilt, Netherlands, Mededel. Verhandel. **69.**
2. ROYAL MAGNETIC METEOROLOGICAL OBSERVATORY, BATAVIA. 1940. Observations made at Secondary Stations in the Netherlands Indies. Climatological Tables. **XIXA.**
3. MCCREARY, F. E. JR. 1959. A Christmas Island Climatological Study. Joint Task Force Seven Meteorological Center. Pearl Harbor, Hawaii.
4. TEWELES, S., L. ROTHENBERG & F. G. FINGER. 1960. The circulation at the 10 millibar constant pressure surface over North America and adjacent ocean areas—July 1957 through July 1958. Mon. Weather Rev. **88:** 137.
5. BERLAGE, H. P. & H. J. DE BOER. 1960. On the southern oscillation, its way of operation and how it affects pressure patterns in the high latitudes. Geofis. pura e appl. **46:** 329–351.

MARINE TRANSGRESSION SEQUENCES IN THE ENGLISH FENLANDS

E. H. Willis

Botany School, Cambridge University, Cambridge, England

The problem of eustasy has produced many theories and wild speculations. These have produced figures from a few millimetres to thousands of feet for the lowering of the sea level during glacial periods to provide water for the vastly increased ice sheets. Flint (1947) has brought the more rational predictions together, particularly those of Antevs (1928) and Daly (1934). By assessing the volume of water locked up in the world's ice sheets during the last glaciation and the degree to which they have now shrunk, Antevs reached the figure of 90 m. and Daly a figure of 85 m. as the level by which the sea has risen since the glacial maximum.

The rate at which the eustatic rise followed the general climatic amelioration has remained a matter of conjecture until relatively recently, almost entirely because of the lack of data. Of recent years, samples dated by the radiocarbon method have begun to make the pattern of events much more clear, although there remains much work to be done. The land/sea level problem is complicated by both the isostatic uplift of the land following the removal of the ice burden from formerly glaciated areas and the long term tectonic movements of the earth's crust. This makes the net rise in sea level often difficult to assess. Nevertheless a number of authors have collected data on the subject (Shepherd and Suess, 1957; Godwin *et al.*, 1958; Graul, 1960; and Fairbridge, 1958, 1961) and in most cases there is a fair measure of agreement. In most instances samples have been obtained with due regard to the assessment of the absolute rise of the oceans since the deposition of the sample and the relation of the position of the sample to its contemporary sea level. It is almost impossible however, when compaction and tidal range are also considered, to obtain a completely unambiguous sample in this connection, and inferences can be drawn, as in many other dating problems, only from a large number of samples.

During the period of the greatest rise in eustatic ocean level, namely in early postglacial times, the corrections might be relatively much smaller than the rise itself, and one might assume that the measure of agreement referred to may be due to this factor. As the rise retarded, however, when the ice sheets approached their new equilibrium limits, the differential movement of the land itself might easily prove greater than that of the sea alone and, as a consequence, inconsistencies between samples might be expected to become more numerous. This makes the timing and extent of marine transgressions that can be ascribed to variations of the eustatic level alone extremely problematical.

The British Isles, covered only in part by the last glaciation, contain areas subject to isostatic uplift and others subject to downwarping. The criterion for describing an area as one of uplift may be taken as the existence of postglacial raised beaches, but that for downwarping is more difficult to establish

since the existence of marine transgressions could be due solely to eustatic influences. If such transgressions occur however, at times when comparable low lying areas suffer no such effect, then the implication is that such an area is one of downwarping. FIGURE 1 illustrates areas of the British Isles known to fall into these two categories and their relationship with similar areas in North West Europe. Isostatic uplift is much in evidence in an area centered upon the Scottish Highlands (Wright, 1937). The Fenlands, on the other hand, form part of a basin of depression centered upon the southern North

FIGURE 1. The principal areas of uplift and downwarping of the post-Glacial period affecting the British Isles and the proximal parts of the Continent.

Sea, for which there is ample evidence of long-term tectonic movement (Stamp, 1936; Godwin, 1945; Pannekoek, 1954; and Van Voorthuysen, 1954). The southwestern part of England, however, by comparison would appear to be free from both these influences. This area may accordingly be used as a control to make a relative assessment of the values of the eustatic and isostatic components of land-sea level changes in other areas. The purpose of this paper is to apply such an assessment to the Fenlands and to establish the chronology of marine transgression sequences in this area.

The Fenlands are characterized by a flat, featureless topography as a result of former marine transgressions. The rich peaty soil produces heavy crops, and it is therefore of considerable economic importance to maintain an efficient drainage system. In the absence of this artificial drainage, the natural drainage

would be poor, as water from the surrounding high ground would enter a river system with an almost negligible gradient towards the sea. The natural landscape therefore would be wood, fen, and open stretches of water only slightly above sea level. The artificial drainage has caused considerable shrinkage of the peat, and rivers at the present time run between artificial banks as much as 20 feet above the surrounding farm land. Ancient water-courses have been unable to shrink because of the high mineral content of the silt, and they are today in evidence as raised ridges of silt meandering across the shrunken peat soils (FIGURE 5).

The natural history and stratigraphy of the Fenlands have been described in some detail by Godwin (1940) and by Godwin and Clifford (1938). Godwin (1948) has also described the Somerset area in similar detail, and since this area is to be taken as the control for the Fenland sequences, it is pertinent to

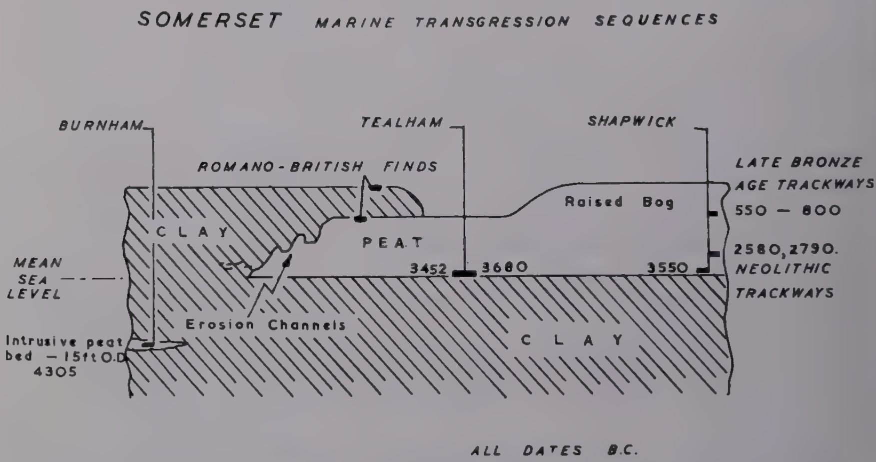


FIGURE 2. Marine transgression sequences in the relatively stable area of Somerset.

review the Somerset sequence first. The area in question is a valley between two low ridges, the Wedmore Ridge and the Polden Hills. With the rising eustatic level of the sea, the valley became an estuary, and marine clay was deposited in the process of the transgression (FIGURE 2). The sea would appear either to have stood still or to have regressed after this episode, for the normally bad drainage of the area gave rise to peat formation, and samples of this peat from its junction with the clay give ages of about 3500 B.C. in 3 instances. If stability is assumed for the reasons previously given, this peat/clay interface must represent the maximum of the main ocean eustatic rise following deglaciation. This interface is very nearly at the height of present day ordinance datum. Peat formation thereafter appears to have been continuous, but the speed of growth was greatly affected by precipitation accelerating markedly in wetter periods giving rise to the phenomenon of recurrence surfaces, recognized widely in northwestern European bog stratigraphy. A second marine transgression does appear to have taken place at the beginning of the Christian era, for this peat is overlain in part by a marine clay from below

which and within which Romano-British finds have been recovered (FIGURE 2). The most probable explanation is, therefore, that no eustatic oscillation of the ocean level occurred (whose amplitude was greater than that terminating about 3500 B.C.) until Romano-British times.

The Fenland marine succession has a different chronology prior to the Romano-British period. At 3500 B.C. there is no evidence of a marine transgression comparable with that which was just ending in the Somerset area. The Jurassic floor of the basin was still above sea level, and drainage was for the

FIGURE 3. A trunk of an oak tree, 67 feet long without branching, which grew on the Jurassic floor of the Fenland prior to the first marine transgression.

most part still relatively good. The landscape was dominated by fairly thick forests, the extent of which may be judged by the specimen in FIGURE 3. This trunk of an oak tree was 67 feet long without branching, and must have fallen with hundreds of its contemporaries as a result of the progressive waterlogging of the ground as each part of the Jurassic floor came nearer and nearer to the contemporary sea level. Oaks cannot achieve the proportions of this one without the competition of a tall forest canopy. This particular specimen was found at the base of the peat at Adelaide Bridge (FIGURE 4), a site toward the Fen margin that became waterlogged later than did the more seaward areas. The outermost tree rings of this trunk gave an age of 2535 B.C. The peat that grew in these waterlogged conditions appears to have been overlain by

the clay of a marine transgression. This clay has a maximum thickness o
22 feet nearer the sea but thins and, finally, disappears farther inland (FIGURE
6). At St. Germans (FIGURE 5) the clay is about 12 feet thick, and peat from
just under it was dated at 2730 B.C. At the limit of the marine clay at Wood
Fen (FIGURE 6) the peat gave an age of 2235 B.C., so that the marine trans-
gression must have been in progress at least between these 2 dates. Water-
logging of the ground and subsequent peat formation appears to have preceded

FIGURE 4. A map of the English Fenlands showing the location of sites from which dated samples have been obtained, and also the present river drainage system into the North Sea.

the actual laying down of the marine clay in any one site by about 300 or 400 years. After the maximum of the transgression had been attained, regression appears to have been relatively rapid, and dates for samples from the peat

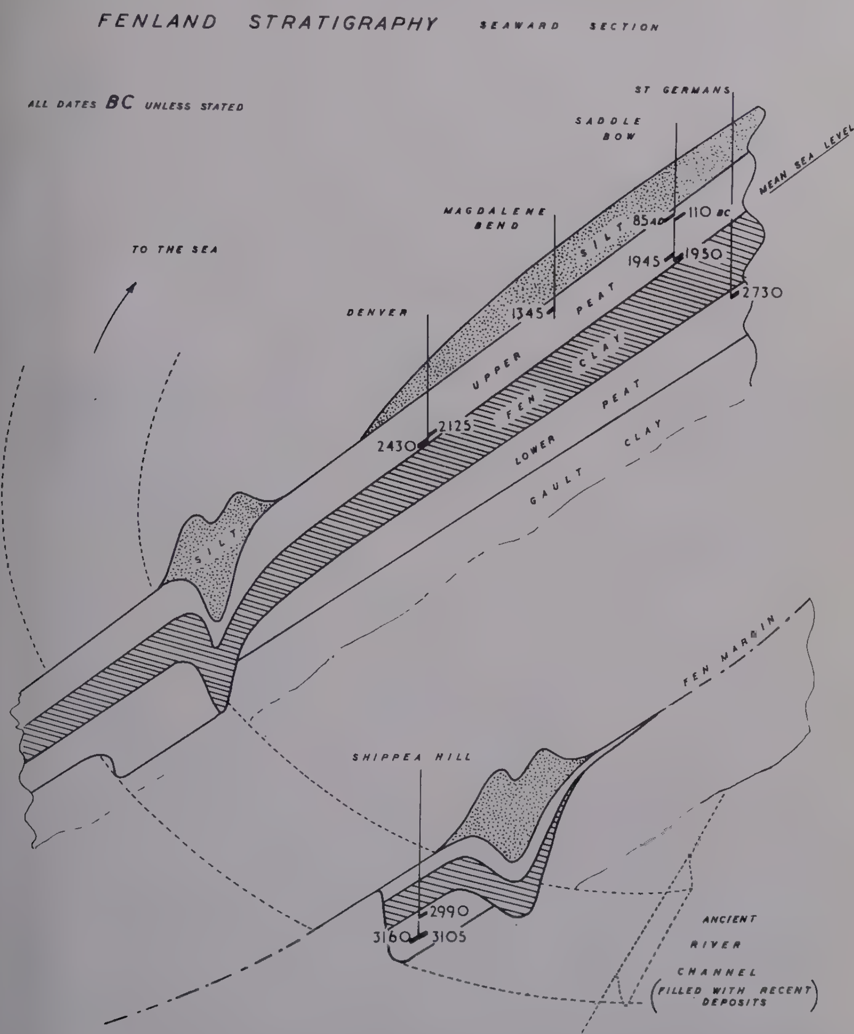


FIGURE 5. The seaward section of the Fenland stratigraphy, showing the age and position relative to the stratigraphy of samples obtained from some of the sites in FIGURE 4. The deposits at Shippea Hill lie in a former river channel.

just above the marine clay vary from about 2200 B.C. inland to about 2000 B.C. nearer the sea. Still being very low-lying, peat formation continued on top of the clay because of continued bad drainage. Bones of a pelican have been found in this context at Saddle Bow, as shown in FIGURE 5 (Forbes *et al.*, 1958). Deposition was relatively slow, however, but was not interrupted by

any further marine influence until Romano-British times when, in the same manner as in Somerset, the peat was overlain by a deposit of silts. Peat from just under this deposit gave ages of 85 A.D. and 110 B.C., so that the conclusion may be reached that the later transgressions in Somerset and the Fenlands are broadly synchronous.

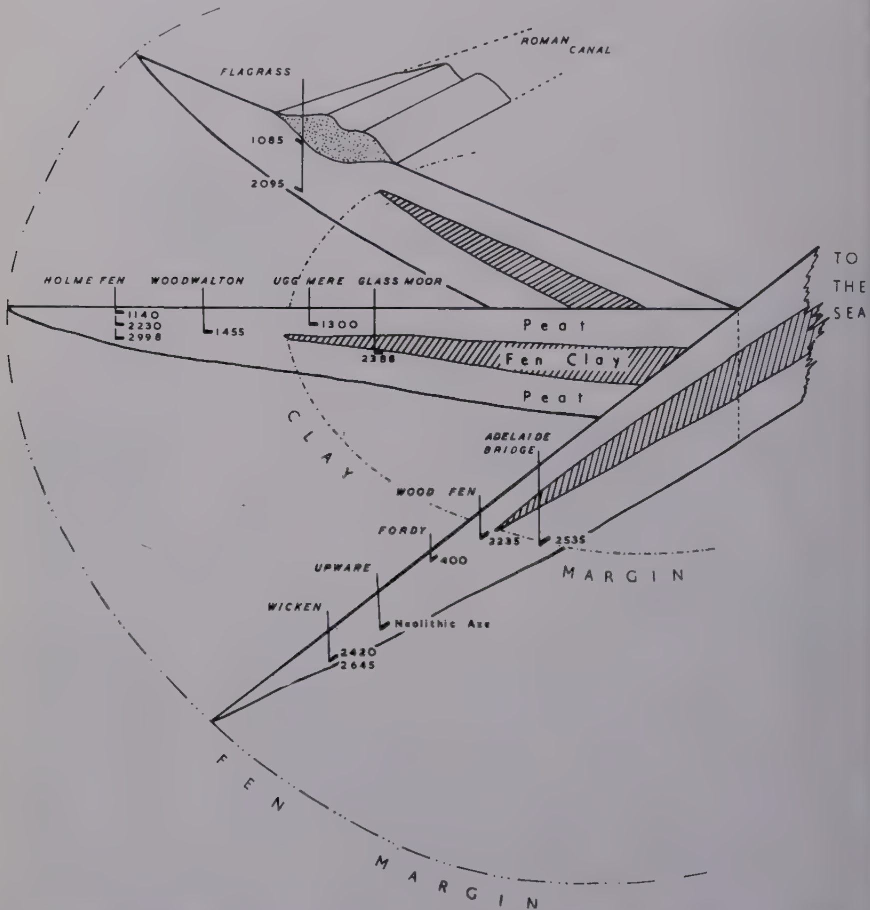


FIGURE 6. The inland continuation of the stratigraphy of FIGURE 5, showing the gradual decrease in thickness of the marine clay. The Roman canal at Flagrass, silted up by the Romano-British marine transgression, is now in evidence as a raised ridge of silt.

Ancient river channels might be expected to develop peat formation earlier than the surrounding ground, and this expectation is confirmed by the excavations at Shippea Hill. Here a sequence of Mesolithic and Neolithic remains have been recovered from peaty deposits in such a channel. The Romano-British transgression was responsible for bringing quantities of silt up the river system and, at Shippea Hill, this silt is found as a ridge referred to previously (FIGURE 5). Other dates from sites shown in the figure are of samples taken

from intermediate points in the stratigraphy, and they amply support the chronology described.

Since it has been argued that the Fenlands are part of a basin of net depression associated with the southern North Sea, it is of obvious importance to compare the chronology thus far described with that of comparable areas along the Dutch coast. The gross lowering of this coast with time was well demonstrated at an early stage of the radiocarbon dating method by de Vries and Barendsen (1954), but later work has revealed the more detailed chronology of phases in this lowering. Pons and Wiggers (1960) investigated the Wieringermeer marine deposits in 3 different localities and have shown that these were laid down at the beginning of the sub-Boreal period and, according to the radiocarbon dates, between 2900 B.C. and 2300 to 2200 B.C. A similar sequence was also found at Meedhuizen-Farmsum (Jelgersma, 1960; Jelgersma and Pannekoek, 1960), and the radiocarbon dates bracket the marine transgression there between 3045 and 2310 B.C. After the clay formation, peat growth started nearly everywhere, often behind the now closed beach-bar coast. The peat growth, unlike that in the Fenlands, was interrupted by other transgressions of shorter duration that laid down the Westfrisian deposits I and II. Since the area may be considered closer to the center of the basin of depression than the Fenlands, the extent of sinking might be greater, and this would be likely to involve further transgressions. A very pronounced Late Roman and early Merovingian transgression is also found in the Dutch sequences and has been dated, as in the Fenlands and Somerset, at about 200 to 500 A.D.

Discussion

It would appear, then, that just after 3000 B.C. a marine transgression took place along the coastal areas associated with the southern North Sea basin, and that a marine regression followed about 2200 B.C. The transgression was most probably brought about by the slow sinking of this basin, since no corresponding transgression is found in the stable area of Somerset. It might have been possible for a positive eustatic component to have aided this transgression, but there is certainly no definite evidence to support this supposition. The cause of the regression might be that the land, after a period of downwarping, rose somewhat and emerged from the sea. It might be more likely, but again is unproved, that a eustatic lowering of the sea level took place to effect the regression. In either case the transgression would be in antiphase with the curve proposed by Fairbridge (1958 and 1961) for the fluctuation of the eustatic level over the past 10,000 years.

The transgression of the Romano-British period appears to have occurred simultaneously in the stable and the unstable areas, and the consequent inference is that its origin is eustatic. This, again, is in contradiction to Fairbridge's argument. It has been shown in this paper that positive evidence for eustatic oscillations is often very difficult to evaluate from the study of marine transgression sequences of the past 5000 years. The land component of land/sea level change might well be less predictable than is commonly supposed, and more work will be needed in other areas before firm conclusions on this aspect can be reached.

Acknowledgment

The stratigraphical evidence in this paper is very largely the work of H. Godwin, F.R.S., and I am extremely grateful for his very considerable help with this problem.

References

ANTEVS, E. 1928. The last glaciation. Am. Geog. Soc. Research. No. **17.**

DALY, R. A. 1934. The Changing World of the Ice Age. Yale Univ. Press. New Haven, Conn.

FAIRBRIDGE, R. W. 1958. Dating the latest movements of the Quaternary sea level. Trans. N. Y. Acad. Sci. **20:** 471.

FAIRBRIDGE, R. W. 1961. Eustatic changes in sea level. Phys. & Chem. of the Earth. Pergamon. **4:** 99–185.

FLINT, R. F. 1947. Glacial Geology and the Pleistocene Epoch. Wiley. New York, N.Y.

FORBES, C. L., K. A. JOYSEY & R. G. WEST. 1958. On post glacial pelicans in Britain. Geol. Mag. **95:** 153.

GODWIN, H. & M. H. CLIFFORD. 1938. Origin and stratigraphy of deposits in the southern Fenland. Phil. Trans. Roy. Soc. London. **229:** 323.

GODWIN, H. 1940. Post Glacial changes of relative land and sea level in the English Fenland. Phil. Trans. Roy Soc. London. **B230:** 239.

GODWIN, H. 1945. Coastal peat beds of the North Sea region as indices of land and sea level changes. New Phytologist. **44:** 29.

GODWIN, H. & A. R. CLAPHAM. 1948. Swamping surfaces in peats of the Somerset levels. Phil Trans. Roy. Soc. London. **B233:** 233.

GODWIN, H., R. P. SUGGATE & E. H. WILLIS. 1958. Radiocarbon dating of the eustatic rise in ocean level. Nature. **181:** 1518.

GODWIN, H. 1960. Radiocarbon dating and Quaternary history in Britain. Proc. Roy. Soc. London. **B153:** 287.

GRAUL, VON H. 1959. Der verlauf des Glazialenstatischen Meirsspiegelanstieges, berechnet an hand von C-14 Datierungen. Deutscher Geographentag Berlin. **20:** 232.

JELGERSMA, S. 1960. Die Palynologische und C^{14} untersuschung einiger Torfprofile aus dem N.S. Profil Meedhuizen-Farmsum. Verhandl. Kon. Ned. Geol. Mijnb. **19:** 25.

JELGERSMA, S. & A. J. PANNEKOEK. 1960. Post-glacial rise of sea-level in the Netherlands. Geol. en Mijnbouw. **39e:** 201.

PANNEKOEK, A. J. 1954. Tertiary and Quaternary subsidence in the Netherlands. Geol. en Mijnbouw. **16e:** 156.

PONS, L. J. & A. J. WIGGERS. 1959. The Holocene genesis of the province of North Holland and the Zuider Zee Region. Tijdschrift Koninklijk Ned. Aardrijkskundig Gen. **77:** 104.

SMITH, A. G. 1958. Post glacial deposits in south Yorkshire and north Lincolnshire. New Phytologist. **57:** 19.

STAMP, D. 1936. The geographical evolution of the North Sea basin. J. Conseil Intern. Exploration de la Mer. **11:** 135.

VAN STRAATEN, L. M. J. V. 1954. Radiocarbon datings and changes of sea level at Velzen, Holland. Geol. Mijnbouw. **16:** 247.

VAN VOORTHUYSEN, J. H. 1954. Crustal movements of the Southern North Sea basin during Pliocene and early Pleistocene times. Geol. en Mijnbouw. **16:** 165.

DE VRIES, H. & G. W. BARENDSEN. 1954. Measurements of age by the radiocarbon technique. Nature. **174:** 1138.

WRIGHT, W. B. 1937. The Quaternary Ice Age. MacMillan. London, England.

THE SEQUENCE OF TERRACES OF THE LOWER THAMES AND THE RADIATION CHRONOLOGY

F. E. Zeuner

Institute of Archaeology, University of London, London, England

Owing to a number of fortunate circumstances, the middle and lower courses of the River Thames provide an exceptionally complete record of the climatic fluctuations of the second half of the Pleistocene. During the first half of this epoch the Thames did not flow through the London area but by-passed it on the north side via St. Albans, its lowermost portion not being known accurately, as the gravels disappear beneath boulder clay east of the Lea Valley. A small pre-Thames river, however, must have existed in the axis of the London Basin at that time, which eventually linked up with the Thames near Staines. There are several events in the history of the river that fix its chronology with a fair amount of accuracy.

Whatever else happened in post-Eocene times, it is generally agreed that by the end of the Tertiary the London Basin was a bay of the sea with beaches that are now found at a height of slightly less than 600 feet on the chalk ridges surrounding the basin, that is, the North Downs on the south and the Chiltern Hills on the north side. The presence of this phase, which would be contemporary with the Calabrian transgression of the Mediterranean, was first recognized by Wooldridge (1927).

Subsequently, the main drainage line developed along the Vale of St. Albans, the history of the phases involved having been published in considerable detail by Wooldridge (1927, 1955, 1958, 1960), Wooldridge and Linton (1955), Hare (1947) and others. The date of the transfer of the river into its present valley is now generally admitted to be virtually contemporary with the second phase of the Antepenultimate Glaciation (Zeuner, 1945, p. 135; Wooldridge, 1958, p. 17; Charlesworth, 1957, vol. II, p. 1017). It is merely a detail whether the ice was directly responsible for this displacement, which is Wooldridge's view, or whether it was a coincidence that at the time of the greatest extension of the ice sheet the pre-Thames river captured the Thames near Staines, the view held by Zeuner. The ice of this glaciation descended into the valley at Hornchurch, (Holmes, 1892; 1894; Wooldridge, 1958, p. 11) and its boulder clay was subsequently covered by the aggradation of the famous Swanscombe Terrace (often called the Boyn Hill Terrace). The age of this aggradation is Great Interglacial, as shown by its connection with the 100-foot sea level, its palaeontological remains, its archaeological content, and its position between the Hornchurch Boulder Clay and the solifluctions of the Penultimate and Last Glaciations. That it cannot be of Last Interglacial age is evident, since it was followed by no fewer than twelve cold phases, suggested by oscillations of the sea-level.

In the area of Dartford, downstream from London, the Thames passes across a barrier of chalk that has had the effect of slowing down the nick-point erosion, stimulated by drops in sea level, in such a way that each drop in sea level is recorded by a separate bench. While it is not possible to recog-

nize consistent benches before the capture of the Thames by the pre-Thames river, there is a clear sequence of 14 benches recognizable in the Dartford and London areas (Zeuner, 1959, p. 360). These have been labeled F to R (including a N′ bench) and the G bench is that covered by the Hornchurch Boulder Clay. This somewhat bewildering number of drops in sea level can be sorted out by some major aggradation phases that follow certain of them. The G bench was followed (after the formation of the Hornchurch Boulder Clay) by the aggradation of the Swanscombe Terrace of Great Interglacial age, grading into an estuarine horizontal surface at 32 m. (100 feet). The N bench carries the Taplow aggradation of Last Interglacial age with a surface at 17 m. (50 feet), while the O bench is covered by the Upper Flood Plain Terrace, the surface of which becomes horizontal at 7.5 m. (25 feet). Finally, the P bench carries a similar aggradation, the Lower Floodplain Terrace, which becomes horizontal at 3 m. above sea level. These estuarine aggradations correspond to the shoreline phases called Tyrrhenian, Main Monastirian, Late Monastirian and Epi-Monastirian by myself (1952, 1953), of which the Taplow and Upper Floodplain Terraces are of Last Interglacial age.

The Upper Floodplain aggradation is followed by the formation of three benches corresponding to three low sea levels, representing three phases of the "Last Glaciation."* Between the phases of Last Glaciation 1 and 2, therefore, there appears a phase of sea level slightly higher than the present that had not previously been recognized but is now amply confirmed by many localities. In south Devon it is covered by 30 and more feet of solifluction material, so that it cannot be mistaken for a Postglacial phase. This is in any case impossible in the Thames, where it is followed by the formation of 2 further benches, the last of which is the one from which the sea level is known to have risen to its present height.

This sequence of numerous drops in sea level punctuated by four recognizable high sea-level phases suggests that many more minor cold phases occurred during the Pleistocene than has been recognized in most areas. There are certain exceptions, however. In particular, the river systems of the periglacial area of central Europe have all revealed similarly complicated sequences (Soergel, 1924, 1925; Grahmann, 1933; Siegert, 1911; Toepfer, 1933; Zaruba, 1957, and others) and so have the glacifluvial terraces of the Alps (Eberl, 1930) and the loesses (for summaries see Zeuner, 1958, 1959, p. 362).

It thus emerges that after the Antepenultimate Glaciation there occurred six cool phases before the main phase of the Penultimate Glaciation. Of these, the last must be correlated with Penultimate Glaciation phase 1, in the sense used by many authors on the continent of Europe. This leaves five cool phases which should have occurred in the course of the Great Interglacial. Similarly, two cool phases are represented by the N′ and O benches, which are followed by three benches corresponding to the Last Glaciation. The N′ bench, however, is a poorly represented bench, whilst the O bench carries the Upper Floodplain Terrace aggradation and thus clearly separates the Main

* These phases are not to be correlated with Wurm 1, 2, and 3 of the Alpine sequence, since Last Glaciation 1 in my nomenclature corresponds almost certainly to the Warthe glaciation of the Scandinavian ice sheet (for details, see Zeuner, 1954).

from the Late Monastirian, the two well known sea-levels of the Last Interglacial.

We therefore arrive at a sequence, *working backwards from the present*, of cool and cold phases separated by certain major warm phases as follows: Last Glaciation 3, Last Glaciation 2, Epi-Monastirian warm phase, Last Glaciation 1, Late Monastirian warm phase, intra-Monastirian cool phase, Main Monastirian warm phase, Penultimate Glaciation 2, Penultimate Glaciation 1, five cool phases during the warm aggradation to the Tyrrhenian sea level of the Great Interglacial, and, two cold phases of the Antepenultimate Glaciation. It is perhaps more than a coincidence that this rhythm corresponds exactly to the periods of low summer radiation as shown by the radiation diagram based on calculations made by Leverrier and Milankovitch (1941), Stockwell and Milankovitch, van Woerkom, and others. Moreover, it agrees for the last six cold phases with the generalized temperature variation of ocean water obtained by Emiliani (1955) by means of the O18/16 ratio. In comparing Emiliani's curve it must be remembered that he was not familiar with the subdivisions of the earlier glaciations, and that his correlation, therefore, must be revised as shown in Zeuner (1959, p. 371). The oscillations, however, remain the same, and the resemblance of the curves is striking.

It thus appears that the evidence of river terraces connected with sea-level oscillations, ocean temperatures, and astronomical evidence for variations in solar radiation are all beginning to present a coherent picture of events during at least the second half of the Pleistocene.

References

Charlesworth, J. K. 1957. The Quaternary Era. 2 vols. Arnold. London, England.

Eberl, B. 1930. Die Eiszeitenfolge im nördlichen Alpenvorlande, 427 pp. Filser. Augsburg.

Emiliani, C. 1955. Pleistocene temperatures. J. Geol. **63**(6): 538–578.

Grahmann, R. 1933. Die Gliederung des Paläolithikums und die Einordung der ältesten Klingenkulturen Deutschlands. Forsch. u. Fortschr. **13:** 465–466.

Hare, F. K. 1947. The Geomorphology of part of the Middle Thames. Proc. Geol. Assoc. **58**(4): 294–339.

Holmes, T. V. 1892. The new railway from Grays Thurrock to Romford. Sections between Upminster and Romford. Quart. J. Geol. Soc. **48:** 365.

Holmes, T. V. 1894. Further Notes on some Sections on the New Railway from Romford to Upminster, and on the Relations of the Thames Valley Beds to the Boulder Clay. Quart. J. Geol. Soc. **50:** 443.

Milankovitch, M. 1941. Kanon der Erdbestrahlung und seine Anwendung auf das Eiszeitenproblem. Ed. spec. Acad. R. Serb. Belgrade. **133**.

Siegert, L. & W. Weissermel. 1911. Das Diluvium zwischen Halle a S. und Weissenfels. Abh. Preuss. geol. Landesanstalt. Berlin (N.f.). **60:** 351 pp., 17 pls.

Soergel, W. 1924. Die diluvialen Terrassen der Ilm und Ihre Bedeutung für die Gliederung des Eiszeitalters. 79 pp. Fischer. Jena, Germany.

Soergel, W. 1925. Die Gliederung und absolute Zeitrechnung des Eiszeitalters. Fortschr. Geol. Palaeontol. Berlin. **13:** 125–251.

Toepfer, V. 1933. Die glazialen und präglazialen Schotterterrassen im mittleren Saaletal. Ber. naturforsch. ges. Freiburg Breisgan. **32:** 1–110.

Wooldridge, S. W. 1927. The pliocene history of the London basin. Proc. Geol. Assoc. **38:** 49–132.

Wooldridge, S. W. 1958. Some aspects of the physiography of the Thames valley in relation to the ice age and early man. Proc. Prehist. Soc. (1957). **23:** 1–19.

Wooldridge, S. W. 1960. The Pleistocene succession in the London basin. Proc. Geol. Assoc. **71**(2): 113–129.

WOOLDRIDGE, S. W. & H. C. K. HENDERSON. 1955. Some aspects of the physiography o the eastern part of the London basin. Trans. Pap. Inst. Brit. Geogr. Publ. No. **21:** 19–31
WOOLDRIDGE, S. W. & D. L. LINTON. 1955. Structure, Surface and Drainage in South east England. 176 pp. Philip. London, England.
ZÁRUBA, Q. Pleistocene terraces of the Vltava river and their relation to a polyglacial sys tem. Actes Congr. Intern. Quat. Madrid-Barcelona. **5**. In press.
ZEUNER, F. E. 1945. The Pleistocene Period. 322 pp. London (Ray Society).
ZEUNER, F. E. 1952. Pleistocene shore-lines. Geol. Rundschau. **40**(1): 39–50.
ZEUNER, F. E. 1953. The three 'Monastirian' sea-levels. Act. 4th Congr. int. Quat. Rome : 1–7.
ZEUNER, F. E. 1954. Riss or Würm? Eiszeit. u. Gegenwart, Öhringen. **4–5:** 98–105.
ZEUNER, F. E. 1958. Dating the Past. 4th edit., 516 pp. Methuen. London, England.
ZEUNER, F. E. 1959. The Pleistocene Period. 2nd. ed., 447 pp. Hutchinson. London, England.

Part V. Cycles and Paleoclimatology

LATITUDINAL PASSAGE: A PRINCIPLE OF SOLAR-TERRESTRIAL CYCLE BEHAVIOR

Leonard W. Wing
Foundation for the Study of Cycles, Ann Arbor, Mich.

Introduction

Much has been written and said about the influence of latitude upon climate and other terrestrial happenings; less about the influence of latitude upon sunspots and other solar happenings. Few investigations if any seem to have attempted to study latitudinal relations from the viewpoint of cycles and spheres: that is, comparative behavior on the sun and earth by latitude.

My own work seems to indicate that some aspects of solar and terrestrial cycle behavior do have angular relationships. It led me to describe earlier, as a postulate, what I now believe is a basic principle of solar-terrestrial cycle behavior. That being the case, it seems possible to take advantage of this in studying solar events. On earth, for example, we can pinpoint latitude more closely than on the distant sun. Because of the seemingly synchronous behavior, it seems possible in some cases to work backwards from latitudes on earth to similar ones on the sun. In this connection, I think that we may be dealing with but a facet of a vast and monstrous field, perhaps that of true ultralong waves; the general outline of this concept I have given in the course of several papers. The possibility that some of our solar-terrestrial cycles may be related to some sort of cosmic waves of extrasolar origin, I have considered already (Wing, 1959). This paper will therefore deal with solar-terrestrial behavior itself.

It has been known for about one century that sunspots after an absence or near absence appear once again in low-middle latitudes, roughly 40° either side of the solar equator. Sunspots appearing later have more equatorward latitudes, the last ones of the interval having the lowest latitude. This brings us to one of the points of this report, "Are these shifts in latitude independent events or do they mark some ripplelike waves passing over the solar surface from poles to equator, invisible or inoperative poleward of the sunspot zone though they may be?"

Our knowledge of behavior on earth, seemingly simultaneous with apparently similar behavior on the sun, may throw some light upon this. The presence on earth of a corresponding behavior clearly a phenomenon of latitude, does indeed suggest ripplelike activity passing equatorward. This I have termed latitudinal passage. The principle here evolved is, I think, but part of a greater phenomenon with which I hope to deal again at another time.

Latitudinal passage occurs in all cycle wave lengths thus far tested for it. It occurs in a wide variety of terrestrial phenomenons—in all thus far tested for it. It suggests very strongly that latitudinal passage may well be a characteristic of any and all solar-terrestrial cycles, irrespective of wave length.

A corollary behavior appears in the higher terrestrial atmosphere, in a few

tests, also seemingly in the higher solar reaches. This I call counter passage. It seems to be a simultaneous, ripplelike behavior moving with a timing reciprocal to that of the equatorward passage at the surface. Presumably, every cycle having equatorward passage at the surface has also counter passage at high altitudes.

I will follow certain conventions in this paper. Time will read left to right; north will always be at the top of all charts and graphs and, conversely, south will always be at the bottom.

Cycles and Pygmatics

The record of sunspots and/or auroras since 649 B.C., when run out on a time chart, gives a measure of 11.084 years, which rounds off to 11.08-year wave length (FIGURE 1). I rather think that this is within 0.01 of being precise, in light of the data available. There is evidence, however, that the manifested length itself has varied longer and shorter at 200-year intervals under the influence of other cycles (Dewey, 1960). That there are many cycles of other wave lengths present in the sunspot time series seems clear. The 11-year manifest cycle seems to be the expression of this dominant, 11.08-year wave length pulled from true expression by many other cycles and influences. This report will touch upon several cycles measured with considerable precision.

I conceive of the 11.08-year wave length as dominant in the sun for reasons that I call pygmatic ones. However, save for influence such as upon auroras and terrestrial magnetism, it is pretty much run-of-the-mill on earth. This brings up an important point. Just because a cycle is overwhelmingly dominant in one event is no reason whatever to expect it to be dominant or even influential in other things. The 9.60-year cycle, for example, is dominant in the ebb and flow of numbers of snowshoe rabbits and lynx in northern North America. In the lives of these animals, it is as overwhelmingly dominant as is the 11.08-year wave length in sunspots—if anything, even more so. It has repeated faithfully since first made known to white man by northern Indians more than 2 centuries ago. A cycle tentatively measured at 3.86 years is of transcendent importance in lemmings and other small rodents of the northern parts of the Northern Hemisphere. The ruffed grouse and its allies respond to a cycle of about 9.93-years wave length over much of the same region where the snowshoe rabbit and lynx respond to a 9.60-year cycle and the lemmings to a 3.86-year one. The response of each to the respective wave length certainly is stronger than that of sunspots to the 11.08-year wave length. The 9.60-year wave length and 9.93-year wave length are commonly called "10-year cycles" and confused with each other, perhaps also with still others. Their epochs pull apart by only about 3 years in a century, which no one in the field would detect. But these cycles are pretty much run-of-the-mill cycles in other events in the same region.

All of these cycles appear to be present in meteorological records: rainfall and runoff, temperature, and barometric pressure. They are of low amplitude and can be detected only in a long time series where each contributes its own little bit in its own little way and in its own little time. Yet none contributes very greatly to the whole. How many cycles there may be is conjectural. It

could be scores, and it could be an infinite number. We have no way of knowing other than that there are many: even in sunspots.

However, just why a phenomenon such as sunspot or rabbit numbers, should respond so overwhelmingly to a particular wave length while rather ignoring others is a puzzle. By appropriate techniques for neutralizing the overwhelming one, we can detect others in many time series. I call this behavior pygmatics in recognition of its parallel to light falling upon plants (Wing, 1956). For reasons of pigment, the leaves overwhelmingly throw back to our eyes the green light waves. We know that all light waves from the sky do fall upon the leaves. Yet the leaves respond chiefly to the green. A strawberry responds to red, the gentian to blue, and the dandelion to yellow. Nevertheless we know that the same mixture of light waves has fallen upon the strawberry as upon the dandelion, and upon the dandelion as upon the gentian. By appropriate filters, we can detect other light waves in the strawberry or in the leaves. In an analogous wave, we can detect by appropriate techniques other cycle wave lengths in a time series, such as that of the sunspots.

We may conceive of the many cycle wave lengths falling alike upon the grouse, rabbit, lemming, and the sunspot itself, just as the many light waves fall alike upon the strawberry, gentian, dandelion, and leaf. For pygmatic reasons, one responds overwhelmingly to 9.93 years, another to 9.60 years, another to 3.86 years, and another to 11.08 years. The mechanism controlling this state of affairs may be ecological, physiological, physical, chemical, structural, electric, magnetic, or whatever. The general term for it would be pygmatics.

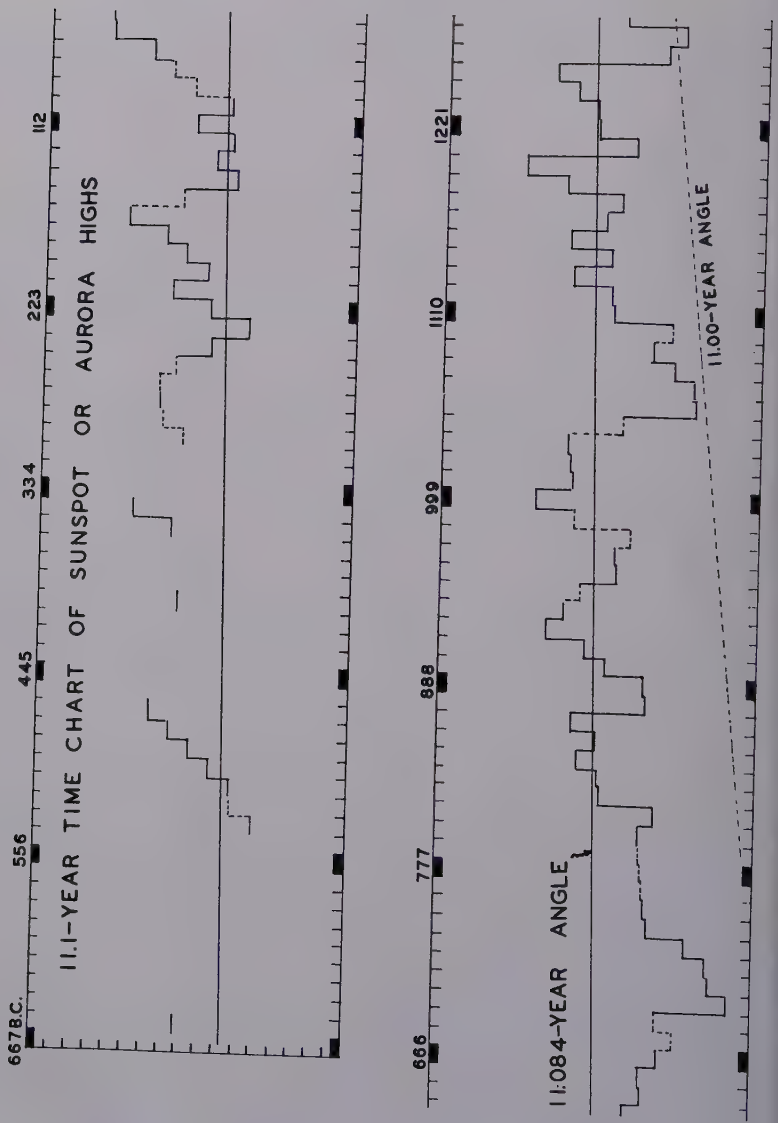
In so far as I know now, every wave length rigorously measured and characterized in terrestrial data, whether measured in days or years wave length, seems to occur also in sunspot data. For all practical purposes, the relative sunspot numbers seem as suitable as any time series.

Latitudinal Behavior

Some years ago the thought occurred to me to test terrestrial data for latitudinal behavior, if any. The tests seemed to suggest a pattern on earth similar to the so-called butterfly pattern of sunspots (Wing, 1954, 1958). I think that most people are aware of the wing-shaped design formed when sunspots are graphed by latitude with respect to time. Earlier spots appear at low middle latitudes on either side of the solar equator. The last ones of an interval appear nearest the equator.

Terrestrial latitudes may be pinpointed closely for one coordinate of an epoch chart. While difficult and time-consuming, the ideal or typical timing of epoch, either high or low, can be measured to give the other coordinate (a physicist might call timing "phase angle"). These allow plotting latitude with respect to time, the resulting graph being the epoch chart.

I have several wave lengths measured with enough precision to run such tests of time and latitude. Those of 9.60 years and 4.222 years from terrestrial sources, 11.08 years from solar sources, and 4.4635 quarter-years, 3.6296 quarter-years, and 5.115 days from extrasolar sources are used here. Time and labor have precluded complete tests with other wave lengths; it has also


11.1-YEAR TIME CHART OF SUNSPOT OR AURORA HIGHS
667 B.C.
556
445
334
223
112
11.084-YEAR ANGLE
11.00-YEAR ANGLE
666
777
888
999
1110
1221

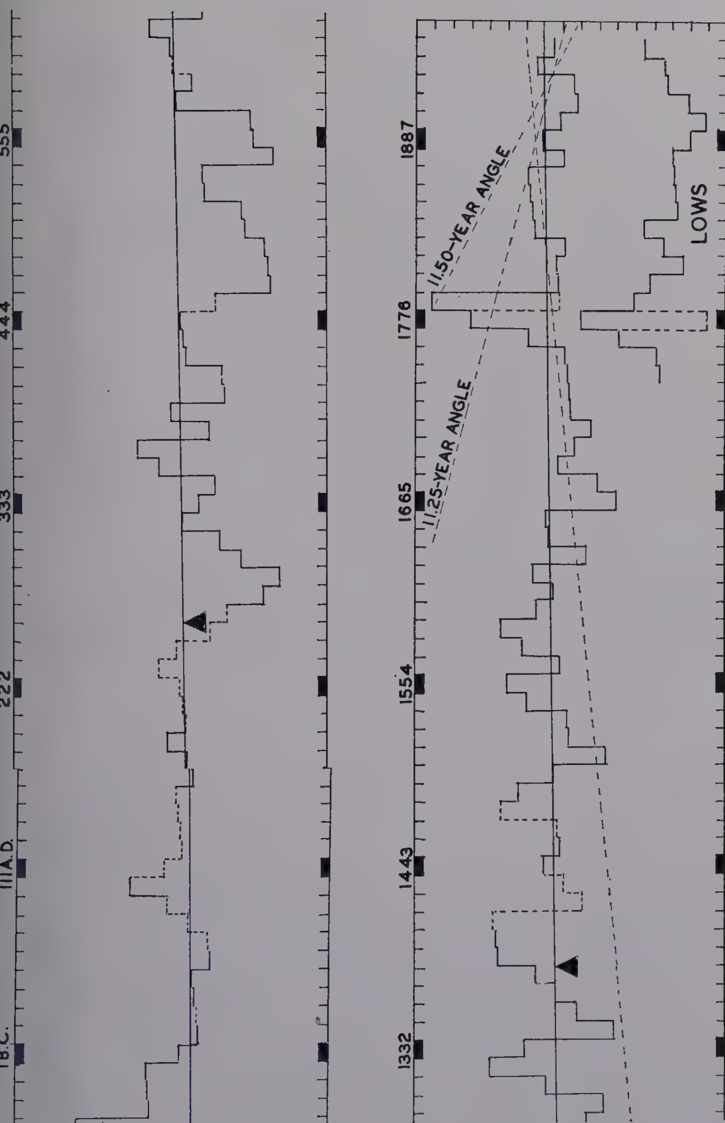


FIGURE 1. An 11.1-year time chart of the sunspot and/or aurora highs since 649 B.C. The median line of the band rises 1.6 years in the 101 intervals between the mid-points (marked by triangles). This rise of 0.016 years for each interval gives the wave length as 11.084 years, rounding off to 11.08 years. The ideal or typical time of high will be 1948.68. The angles taken by any wave length of 11.00 years, 11.25 years, 11.50 years are indicated at the right. Reproduced by permission of the *Journal of Cycle Research* (Wing, 1957).

precluded tests with these same wave lengths in a host of other widespread events. Initial work, however, suggests the same pattern for all.

I have determined, by periodic tables in most cases, the ideal or typical timing for the several wave lengths in the various phenomena. With few exceptions, the tables use deviations of logarithms from moving averages. High has been determined for the 11.08-year wave length in the sunspot-aurora series and by periodic tables of the relative numbers for other wave lengths. Terrestrial data will include some from the ionosphere (foF2, hF) and some from the surface (H-magnetism, barometric pressure, temperature, precipitation, animal numbers). Precipitation will include a few long records of river runoff and lake levels, and one long record of continental varves. Presumably these reflect rainfall, though usually with a lag. I have used tree rings variously in the past, but I have not computed as yet enough epochs distributed widely in latitude, wholly because of the limits of labor.

Epochs of High and Low

A point that would seem explainable some day after enough research results have accumulated will occur to some in the course of reading this paper. It appears that although all wave lengths here tested in terrestrial data occur also in sunspots, phases are not always respectively alike. My meaning will become clear by an example. The timings in barometric pressure with respect to the 3.6296 and 4.4635 quarter-year wave lengths are of low epoch in pressure and of high epoch in sunspots. For the 5.115-day wave length, on the other hand, the timings are of high epoch in pressure and of high epoch in sunspots. Here we have a case of epochs alike in sunspots (that is, highs) but dissimilar in barometric pressure (that is, lows for 2 wave lengths, highs for the third).

Cycles and Record Length

On the basis of my own experience, I feel the need of a few comments on data and their use. In the first place, records short with respect to cycle wave length have a limited use. A few repetitions of even a dominant cycle may have the timing pulled one way or the other from typical because of other cycles or because of random happenings. This may distort the impression of the real conditions. I doubt if fewer than 10 or 12 repetitions of a cycle will give a good idea of behavior. However because data are generally hard to obtain at best, one must work with what can be obtained within reason. Studies with shorter wave lengths are therefore more likely to be productive than studies with longer ones. The relative length of record will be greater. Another point is the matter of techniques, which is of paramount importance. Subjective studies may lead one astray rather easily. Hence objective techniques are essential. Unfortunately, there is not the great body of methods for cycle analysis as for statistics. Statistical procedures themselves are rather limited in use as methods of cycle analysis.

The Dominant Cycle in Sunspots

The first wave length that I should like to examine is the 11-year dominant in sunspots. As said earlier, I have taken the long record that is relatively

complete since 294 B.C. but fragmentary from 694 B.C. to 294 B.C. (Schove, 1956). While this record is based upon both sunspot and aurora maximums (presumably also plus a bit of subjective decision), it appears that the timings of the two are in substantial agreement. It will be shown later in this paper why this may be so. Be that as it may and rough though the records of this series may be, they seem to be about the best that can be had at this time for the study of the dominant wave length in sunspots.

An 11.1-year time chart of the highs gives an opportunity to measure the wave length with some closeness (Wing, 1957). There are 203 intervals of 11.1 years each from 294 B.C. at position 6.7 in the 300.7 B.C. interval to the high of 1949, a span of 2,243 years (FIGURE 1). This length of the record is more than 200 times the wave length itself. We may divide the period into two equal parts of 102 intervals each, the 821.4 A.D. interval being overlapped and common to both parts. There are therefore 101 intervals between the midpoints of the respective 102-interval parts.

A rise to the right by a band characterizes the situation when the wave length of a cycle is *shorter* than the interval of the time chart itself. Hence, the rise per interval subtracted from the length of the interval itself gives a measure of the wave length. In case of something like the sunspots, this is the best measure that can be obtained, although closer measure may be possible sometimes with other wave lengths in other events. The median of the first part falls at position 8.3 after base in the 255.3 to 266.4 A.D. intervals as indicated by a triangle in FIGURE 1. The median of the second 102 intervals falls at position 6.7 after base in the 1376.4 to 1387.5 A.D. interval and is indicated also by a triangle. The difference of 1.6 positions (years) between the respective medians indicates a rise of the band itself by 1.6 years in the 101 intervals between the mid-points, which would be 0.016 years per interval. This adjustment to the 11.1-year interval of the time chart gives a measure of 11.084 years for the wave length of the cycle, which rounds off to 11.08 years. I believe that 11.084 years is probably within 0.010 years of correct as we can best measure now. In other words, the dominant cycle wave length in the sunspots lies between 11.074 years and 11.094 years (Wing, 1957; Dewey, 1958, 1960).

Various investigators have reported in the past various wave lengths for the 11-year manifest cycle in sunspots. I have indicated at the right end of the time chart the angles that some of these might make with respect to the time chart. Eleven and one half years and $11\frac{1}{4}$ years are clearly too long, they depart so hurriedly from the band of highs. Eleven years is closer but so short as to leave the time chart before reaching backward as far as the middle; 11.1 years is too long, as may be seen by the rise of the band itself in the time chart.

The next point to consider is what solar latitude ideal time of high might represent. I have made a conjecture of midway between the solar equator and 40° solar latitude, which gives 20° latitude either side of the solar equator as that of typical or ideal sunspot high. Presumably, it would be the same latitude for ideal of typical sunspot low. A preliminary check of recent highs indicates that this may be a bit too far poleward. The median solar latitude at high during the past 80 years falls at 19.1° solar latitude, the mean at 17.9°.

Whatever the exact latitude sunspot high may represent, however, any variation from precise will not change conclusions on the principle of latitudinal passage itself.

FIGURE 2 is an epoch chart of the sunspots and latitude with respect to time. The chart allows for full pole to equator latitude. I have diagrammed the lines of passage for an ideal wave of 11.08 years with timing of 1948.68 as of 20° latitude (Wing, 1957), the timing being from the time chart just discussed (FIGURE 1). The passage rate is computed as 90° passage = $\sqrt{\frac{1}{2}\,(\text{wave length})^2}$, the ratio of passage to wave length indicated by previous studies (Wing, 1954, 1957). This same ratio, it may be noted, will be used for wave lengths additional to that of 11.08 years.

The 11.08-Year Cycle and Solar Prominences

From the works of various writers, I gain the impression that (1) some solar prominences seem associated with sunspot zones, (2) others tend to be most numerous when sunspots are most numerous, and (3) those of high solar latitudes appear farthest poleward when sunspots are most numerous. The last point suggests behavior analogous to that of the sunspots themselves but in reverse. I have added reports of prominences to the previous diagram and also the reciprocal passage line for an ideal wave length of 11.08 years, the timing now being as of 1948.68 at 70° latitude (FIGURE 3). The line fits surprisingly well the behavior of the higher latitude prominences. It suggests very much that in some way a reciprocal behavior does indeed occur and that it involves the prominences of higher latitudes (which are also of higher altitudes).

The 11.08-Year Cycle and Latitudinal Passage

The immediate point of this is not wholly interest in behavior on the sun. If latitudinal passage occurs across the earth equatorward, and if the sun has both a passage equatorward and another passage poleward, it does not take much scientific imagination to wonder if by analogy we might suspect a poleward phenomenon also at high terrestrial altitudes (that is, counter passage). For every wave length, there might be both an equatorward passage at the surface and a high altitude, poleward counter passage. Later, evidence will be presented for this thesis.

In a sense, we may measure a cycle and determine its passage (and sometimes counter passage) on earth and, with some justification and hope, apply the results to the sun. Here on earth we can trace a cycle latitude by latitude through the rise and fall in the fortunes of terrestrial phenomena, these being the instruments, gauges, counters, and dials that register the passing of the cycle. One might assume that other planets behave similarly. This we could ascertain by assembling and studying the proper data.

I shall test terrestrial data for latitudinal passage with respect to the 11.08-year wave length, using the typical timing of 1948.68 as of 20° solar and terrestrial latitudes. Initially, I took from the Claytons' *World Weather Records* all the precipitation reports 10 times the wave length; that is, beginning in 1841 and continuing through 1950. In order to cover more latitudes, these

FIGURE 2. Pole to pole epoch chart showing the wing-shaped or butterfly design formed by sunspots. The manifest times of high are indicated by letters posted at 20° latitude. The passage line is that for an 11.08-year wave length posted as of 20° latitude with timing of 1948. 68 from FIGURE 1. Reproduced by permission of the *Journal of Cycle Research* (Wing, 1957).

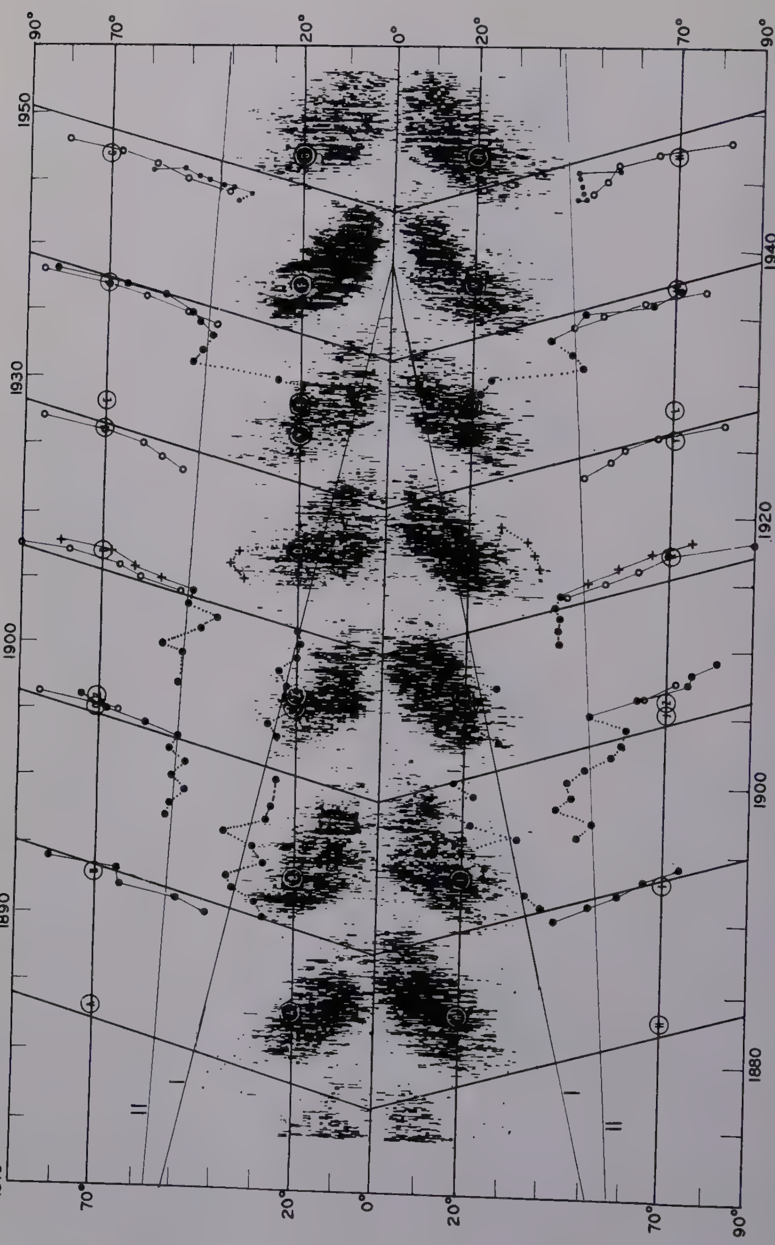


FIGURE 3. To FIGURE 2 has been added solar prominences from several sources and a reciprocal passage line. The high latitude solar prominences fall rather well along the reciprocal passage line. The manifest times of high have been entered by letters at 20° and 70° latitude. Reproduced by permission of the *Journal of Cycle Research* (Wing, 1957).

have been supplemented by stations in other latitudes but for fewer years, which gives a total of 39 (TABLE 1). Additionally, some rivers, a lake, and a continental varve record are included, runoff reflecting rainfall, albeit with a time lag.

If the thesis of latitudinal passage is correct, any passage behavior on earth should be synchronous with that on the sun and should give a wing-shaped pattern, poles to the equator, like the wing-shaped pattern known for sunspots. Times of epoch should appear earlier in the higher latitudes and later in the lower ones. Timing with respect to the 11.08-year wave length should agree with the timing of the 11.08-year wave length in sunspots as of about 20° latitude. This presumes that solar-terrestrial relationships are angular, which seems to be the case. FIGURE 4 tests this thesis. It plots by latitude and calendar the timings of epoch with respect to the 11.08-year wave length in precipitation at the 38 stations. Data for six rivers, one lake, and one set of continental varves makes a total of 46 stations in all (TABLE 1). They cover latitudes from Jacobshavn, Greenland (69.2° N), to Wellington, New Zealand (41.3° S). As usual with man-kept records, the longer ones are in the mid-latitudes, the zone where live people with most interest in recording natural events.

The earliest epochs are farthest poleward; they become progressively later equatorward. The band of epochs tends to center rather well upon the presumed passage line for a wave length of 11.08 years in the sunspots with timing of 1948.68, both the wave length and timing in FIGURE 4 being derived from FIGURE 1. It should be emphasized that the passage line is simply one presumed to be in the sunspots moved bodily from the solar to the terrestrial sphere (90° passage $= \sqrt{\tfrac{1}{2}\,(\text{wave length})^2}$). The timing is as of 20° solar latitude transferred also to 20° terrestrial latitude. Neither the passage nor the timing is based upon precipitation. It should be emphasized also that although the 11.08-year cycle appears to be active in precipitation, its amplitude is low, about that of many other cycles.

Why the epochs should switch from highs to lows rather than being all highs or all lows I do not know. The problem appears to have some sort of geographic explanation, for parallel behavior appears in the epochs for the 9.60-year wave length in precipitation, using the same terrestrial and solar data but rearranging them with respect to the 9.60-year wave length (FIGURE 5; TABLE 1). It occurs also in tree rings. Previously a biological explanation was supposed, but this may not necessarily be true.

The 9.60-Year Cycle and Latitudinal Passage

The line in FIGURE 5 is once again the presumed passage line for a 9.60-year wave length in sunspots with the ratio of 90° passage $= \sqrt{\tfrac{1}{2}\,(\text{wave length})^2}$, and timing of 1955.65 at 20° solar latitude moved bodily from the solar to the terrestrial sphere. The timing and presumed passage line fit rather well the band of epochs in precipitation. We shall see regularly that this fitting-rather-well is a characteristic of all cycle wave lengths tested in solar and terrestrial phenomenons. It argues strongly that in some way, some ripplelike activity passes over the two spheres simultaneously, poles to equator.

Table 1

Epochs of 11.08- and 9.60-Year Wave Lengths in Precipitation

Station	Years of record	Length (years)	Geographic latitude °	Geomagnetic latitude °	Geographic longitude °	Time of epoch 11.08-year wave length	Time of epoch 9.60-year wave length
Jacobshavn, Greenland	1873–1950	68	69.2 N	79 N	51.0 W	1946.07	1953.72
Haparanda, Sweden	1860–1950	80	65.8 N	63 N	24.2 E	1943.69	1953.80
Stykkisholm, Iceland	1856–1950	85	65.1 N	70 N	22.8 E	1944.10	1953.90
Helsinki, Finland	1845–1950	96	60.2 N	59 N	25.0 E	1948.08	1956.35
Uppsala, Sweden	1851–1950	90	59.8 N	58 N	17.6 E	1947.42	1955.49
Edinburgh, Scotland	1785–1950	156	55.9 N	59 N	3.2 W	1945.32*	1954.37
Copenhagen, Denmark	1821–1950	120	55.7 N	56 N	12.6 E	1946.02*	1954.53*
Greenwich, England	1841–1950	100	51.5 N	54 N	0.0	1943.99*	1952.32*
Frankfurt, West Germany	1837–1950	104	50.1 N	51.5 N	8.7 E	1944.48	1954.92*
Budapest, Hungary	1841–1947	97	47.3 N	47 N	19.0 E	1947.21	1953.78
Lyons, France	1841–1950	100	45.7 N	48 N	4.9 E	1945.82	1952.78
Milan, Italy	1764–1950	177	45.5 N	47 N	9.2 E	1945.29*	1953.35*
St. Paul-Minneapolis, Minnesota	1837–1950	104	45.0 N	56 N	93.0 W	1944.69*	1955.88
Toronto, Canada	1843–1950	98	43.7 N	54 N	79.4 W	1947.82	1954.72
Albany, New York	1826–1950	115	42.6 N	53 N	73.8 W	1946.66	1956.07
Boston, Mass.	1818–1940	113	42.4 N	53 N	71.1 W	1947.89*	1956.11
Rome, Italy†	1782–1950	139	41.9	42 N	12.5 E	1948.31*	1955.78*
New York, N.Y.	1826–1950	115	40.7 N	52 N	74.0 W	1955.90	1955.62
Philadelphia, Pa.‡	1820–1950	117	40.0 N	51 N	75.2 W	1946.69*	1956.52
Portsmouth, Ohio	1830–1950	111	38.7 N	50 N	83.0 W	1945.42*	1957.36
St. Louis, Mo.	1837–1950	104	38.6 N	50 N	90.2 W	1945.24*	1953.60
Santa Fe, N.M.	1850–1952	93	35.7 N	45 N	106.1 W	1946.29*	1956.47*
Charleston, S.C.	1832–1950	109	32.8 N	44 N	79.9 W	1946.99	1956.82*
Jerusalem, Israel	1846–1950	95	31.8 N	29 N	35.2 E	1946.16*	1955.90
Havana, Cuba	1859–1950	82	23.1 N	34 N	82.4 W	1950.60*	1957.37*
Calcutta, India	1829–1950	112	22.5 N	12 N	88.4 E	1948.72*	1958.49
Bombay, India§§	1817–1950	124	18.9 N	9 N	72.9 E	1950.54	1958.94*
Barbados, B.W.I.	1853–1950	88	13.1 N	24 N	59.2 W	1946.68*	1959.66*
Madras, India	1813–1950	128	13.1 N	3 N	80.2 E	1950.96	1956.89*
Bangalore, India	1837–1950	104	13.0 N	4 N	77.6 E	1947.76*	1958.06*
Quixeramobim-Fortaleza, Brazil	1849–1950	92	5.3 S	5 N	39.3 W	1948.42	1957.55*
Batavia, East Indies	1864–1950	77	6.2 S	5 S	106.8 E	1950.88*	1957.52*
Rio de Janeiro, Brazil	1851–1950	90	22.8 S	12 S	43.2 W	1949.98	1955.80*
Sydney, Australia	1840–1950	101	33.9 S	43 S	151.2 E	1949.40*	1953.95*
Cape Town, Union of South Africa	1838–1950	103	33.9 S	33 S	18.5 E	1948.36	1956.15*
Adelaide, Australia	1839–1950	102	34.9 S	48 S	138.6 E	1948.16*	1954.83*
Aukland, New Zealand	1853–1950	88	36.8 S	40 S	174.8 E	1949.74	1956.32
Wellington, New Zealand	1862–1950	79	41.3 S	45 S	174.8 E	1948.91	1956.72*
Lake Vänern, Sweden	1807–1954	137	58.4 N	58 N	12.4 E	1945.87*	1952.68*
Göta River, Sweden§	1808–1957	140	58.4 N	58 N	12.3 E	1945.68*	1953.65*
Nemunas River‖	1812–1943	122	55.1 N	53 N	23.1 E	1945.05*	1953.88
Rhine River	1808–1957	140	47.5 N	48	7.6 E	1947.94	1956.73
Danube River	1838–1957	110	44.7 N	43 N	22.4 E	1949.11*	1955.22
Lake Saki,¶ U.S.S.R.	1683–1894	212	45.2 N	42 N	33.6 E	1945.96*	1953.12*
Nile River, Egypt**	1698–1943	236	0–5 N††	0	32.5 E	1949.39*	1956.80*
U-magnetic value	1871–1946	66	0	0	—	1950.78	1957.75

* Low.
† Record for 1931–1940 missing.
‡ Record for 1872–1875 missing.
§ Water year, October 1 to September 30.
‖ Water year, November 1 to October 31.
¶ Continental varves (Shostakovich, 1934).
** Jarvis (1953).
†† Catchment area.
§§ June.

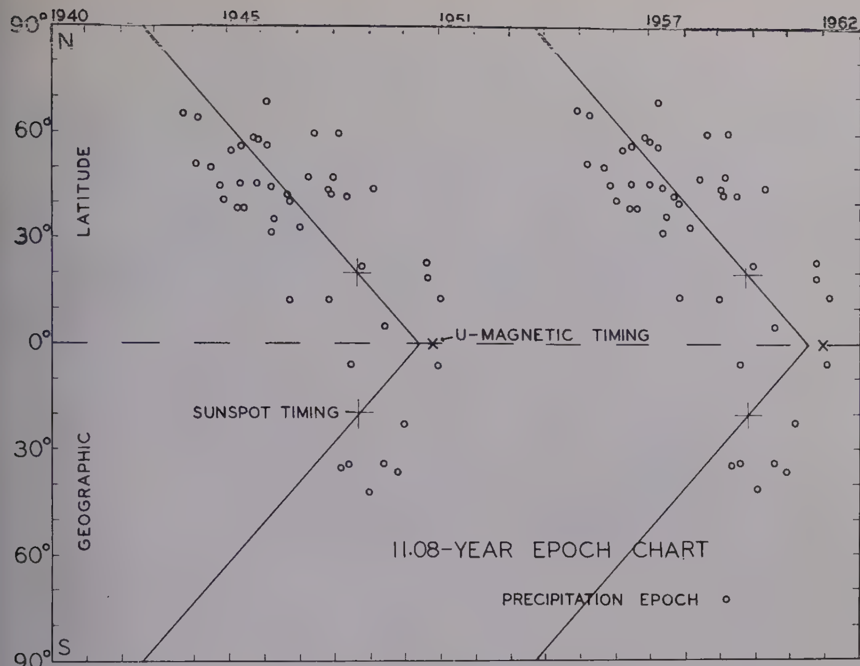


FIGURE 4. An 11.08-year epoch chart of precipitation. The epochs fall into a band near the passage line from FIGURE 2. The poleward epochs tend to be earlier, the equatorward ones later.

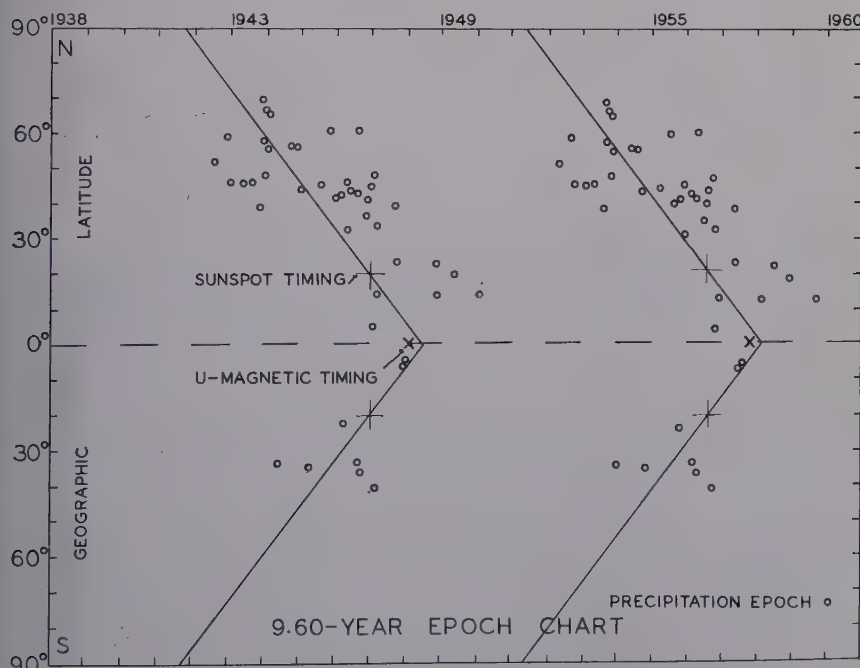


FIGURE 5. A 9.60-year epoch chart of precipitation. The epochs fall into a band near the passage line of the same wave length from sunspots posted as of 20° latitude. The poleward epochs tend to be earlier, the equatorward ones later.

FIGURE 6 is an epoch chart from North America of current timing as highs of past rabbit epochs since 1793 (Wing, 1960). The passage line marks that of a 9.60-year wave length in sunspots, as in FIGURE 5. The letters x and x represent the respective medians north and south of 50° latitude. Each circle marks the current time of high as based upon high or low reports of the past brought down to date by multiples of 9.60 years (with an adjustment of one half the wave length for lows). Unlike the epochs of precipitation, however,

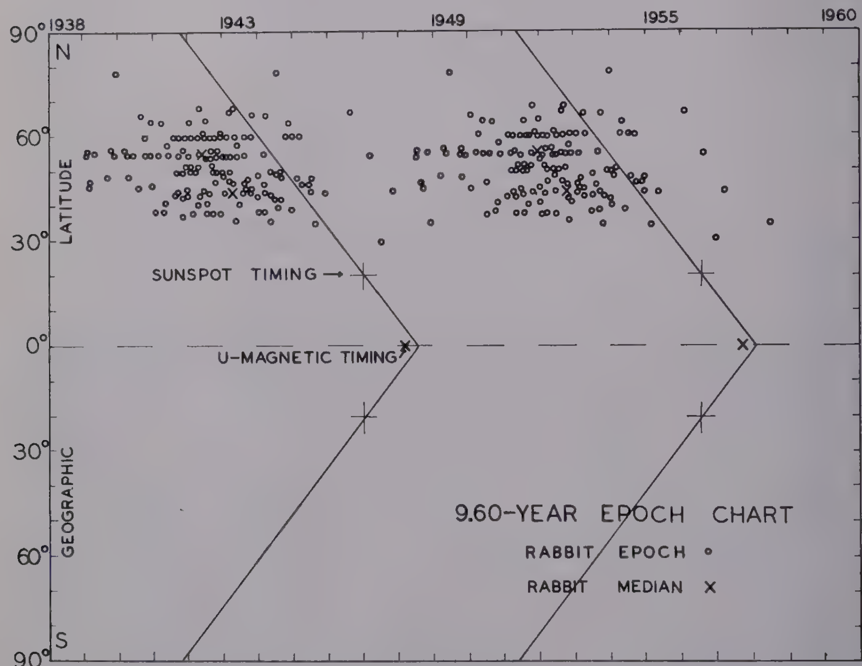


FIGURE 6. A 9.60-year epoch chart of the snowshoe rabbit highs (or lows) brought down as high epoch in current time: X marks the median timing for those highs north of 50° latitude and south of 50° latitude, respectively. The passage line is from FIGURE 5. This is the dominant wave length in the snowshoe rabbit fluctuations and has come true since reported to the white man by the Indians more than two centuries ago.

these are all "highs." (In the present state of cycle research, there is always the danger that highs and lows may be forced into the pattern as in FIGURES 4 and 5. The rabbit epochs are introduced here because this cannot happen, as they are all high epochs.)

An additional reason for introducing FIGURES 6 and 7 here is that the 9.60-year wave length is the dominant cycle in the affairs of rabbits (and lynxes) in northern North America. This has been known by the white man for more than 2 centuries and, doubtless, for milleniums earlier by the Indians. The cycle is so overwhelmingly dominant among northern rabbits that Indians take account of "rabbit years" in their culture. It is therefore a wave length

of terrestrial identification that is being tested in solar and terrestrial affairs. This cycle has been repeating true since its discovery, which is also some sort of record.

FIGURE 7 uses time-chart determinations of timing for the sites having records from a number of successive highs and lows. Such relatively complete records of rabbit abundance and scarcity are few indeed. However, they are valuable from many standpoints, one being a reduction in likely errors. The several time-chart determinations clearly show the apparent latitudinal aspect

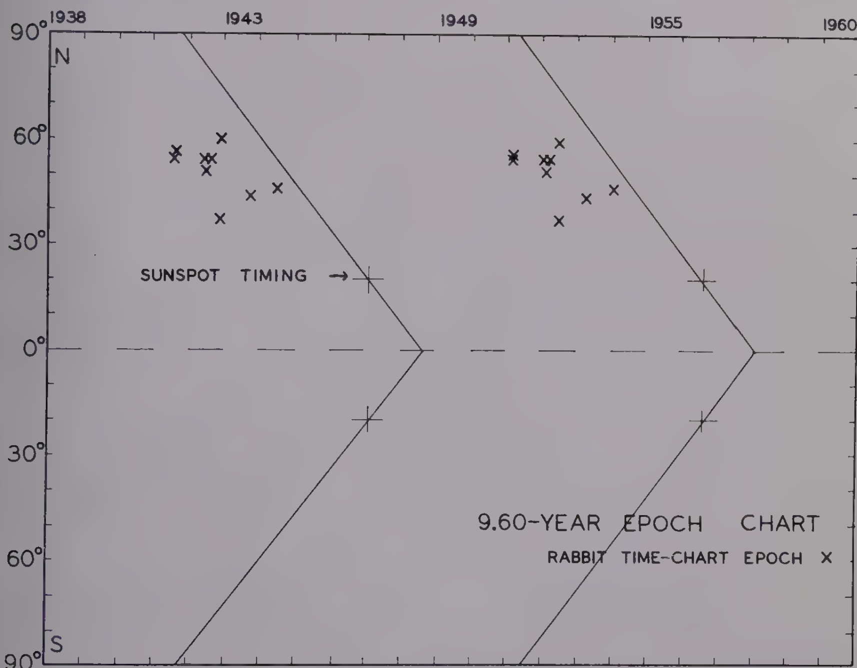


FIGURE 7. A 9.60 epoch chart of snowshoe rabbit time-chart determinations of high epoch. The rabbit timings precede the passage line. It may be that they indicate the wave length better than do the sunspots, the wave length being so dominant in the rabbit.

of rabbit numbers. The fact that the general rabbit band of highs in both FIGURES 6 and 7 precedes that of the sunspot timing may reflect a biological influence. It could be, however, that rabbit reports more correctly define the timing than do the sunspots themselves, the 9.60-year wave length in rabbits being subject to influence by no stronger cycle and perhaps by fewer cycles than are the sunspots. Sunspots are dominated by the 11.08-year wave length; additionally, they seem to be influenced by a host of other cycles, the 9.60-year being but one of many. In the rabbit life history, on the other hand, the 9.60-year cycle is of transcendent importance. Rabbits, by concentrating on this wave length, may in a sense indicate the timing more clearly, as though they were "tuning in" on something of a 9.60-year wave length.

The 4.222-Year Cycle and Latitudinal Passage

Some time ago, I reported the behavior of several wave lengths initially derived from the flight years of the pine grosbeak in the lake states (Wing, 1954, 1956). One of the cycles in this complex measures 4.222-years' wave length. A test in the environment shows that it is present in tree rings and in temperature. Presumably it is present in other environmental events. I might add that it is the only wave length that I have tested in records from the geologic past. It seems to be present in Pleistocene varves and Eocene sediments.

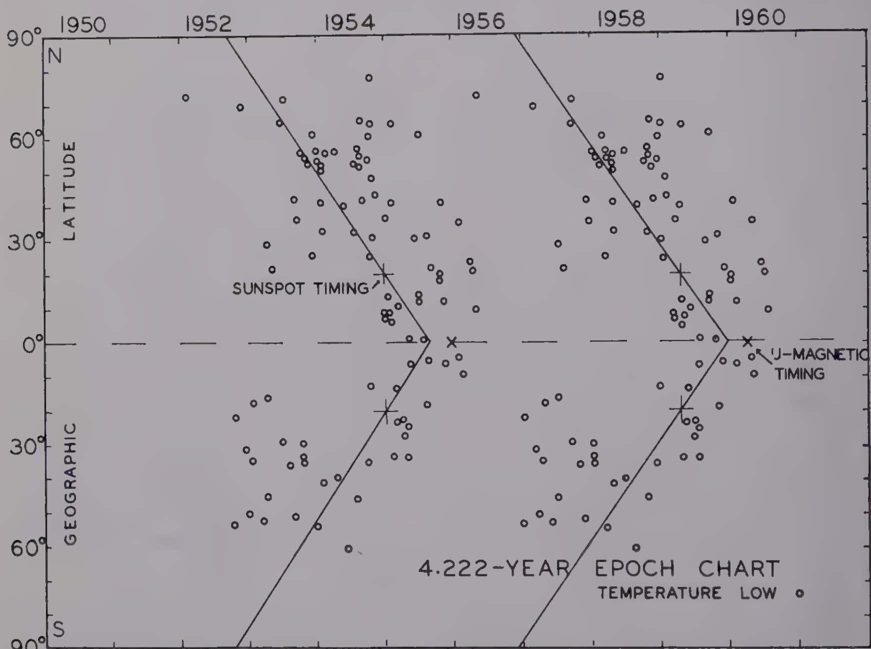


FIGURE 8. A 4.222-year epoch chart of temperature lows by geographic latitude. The lows tend to occur earlier in the more poleward regions, later in the more equatorward ones. Seventy-seven of the 97 lows fall within the half wave length centered about the passage line. The passage line itself is computed from sunspot data.

Its wave length in these records is identical with that of today within the limits of precision enforced by the length of the record. Evidently this wave length has continued unvarying for many millions of years.

FIGURE 8 is an epoch chart of the lows in temperature with respect to the 4.222-year wave length. The band of lows shows a later timing equatorward, just as do the epochs for the 11.08-year and 9.60-year wave lengths in precipitation. It was the behavior of these epochs of temperature on this epoch chart that first suggested the presence on earth of a passage phenomenon synchronous with that on the sun and forming a "butterfly" design. This chart is responsible also for the ratio of $\sqrt{\frac{1}{2}\,(\text{wave length})^2}$ for 90° of latitudinal passage.

All epochs of FIGURE 8 are alike (lows), which offers a chance to test the pattern that might emerge with geomagnetic latitude. FIGURE 8 itself uses geographic latitude, and the clustering about the presumed passage line from the sunspots can be seen readily. FIGURE 9 uses the same epochs but plots them by geomagnetic latitude and time. As a wholly random thing, we should expect the epochs of both FIGURES 8 and 9 to be equally distributed in equal units of calendar time. Thus if we take any unit of calendar time equal to the wave length and divide it into two halves, just as many epochs should fall by chance in one half as in the other. At random, there will be no difference

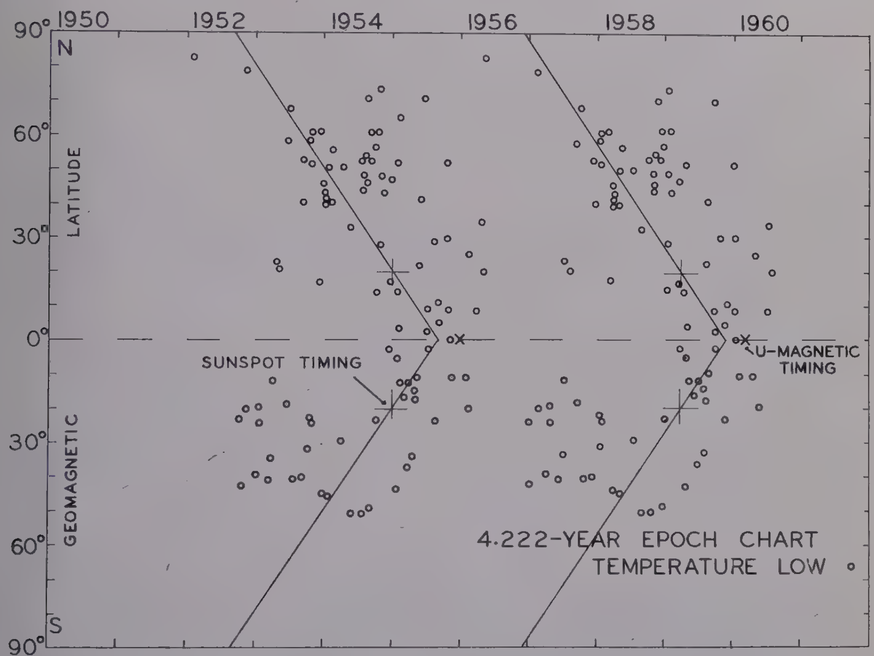


FIGURE 9. A 4.222-year epoch chart of temperature lows by geomagnetic latitude. The band of lows is looser than that in FIGURE 8. Sixty-eight of the 97 lows fall within the half wave length centered upon the passage line.

between one half of the wave length (any 2.111 years) and the other half. Any concentration of epochs will be a measure of nonrandomness.

I have used the presumed passage line as a reference for setting off segments of calendar time equal to the wave length. Seventy-seven of the 97 epochs in FIGURE 8 fall in the half of the wave length centered upon this presumed passage line, which is the reference here, while 20 fall in the other half. Roughly, the epochs are four times as numerous in the centered half as in the other half. At random, they should be equally distributed. In FIGURE 9, which graphs epochs by geomagnetic latitude, 68 fall in the half of the wave length centered upon the reference line as against 29 in the other half. As the epochs have a greater concentration in the centered half of the wave length when considered

by geographic latitude, we must interpret this as indicating a greater association with geographic than with geomagnetic latitude. This indication may turn out to be meaningless, but at the moment it seems worth noting.

The epochs seem satisfactorily distributed in longitude. All 10° zones are represented except 61° to 70° east, 121° to 130° east, 161° to 170° east, 121° to 130° west, and 141° to 150° west. The largest number in any one 10° zone is eight in the 71° to 80° west zone. It hardly seems that lack of wide distribution by either latitude or longitude could account for the difference in concentration of epochs between FIGURES 8 and 9. We seem safe in assuming that it probably reflects something real.

The 3.6296 Quarter-Year and 4.4635 Quarter-Year Wave Lengths

Among the several additional wave lengths tested are two, one shorter than a year and one longer. These are of 3.6296 quarter years (331.42 days) and 4.4635 quarter years (407.56 days), respectively. These wave lengths are of extrasolar identification (Wing, 1959). They have been measured as of quarter years to reduce the work load. Four quarters in a year (January to March, April to June, July to September, October to December) reduces the number of items to one third of those in 12 monthly readings. It does not change anything otherwise, just as measuring something in inches, feet, or yards will not change anything.

There are 15 stations in the New World that form a rough transect from Upernavik, Greenland, (72.8° N) to Laurie Island, Antarctic Ocean (60.7° S), one station for each 10° zone of latitude. There are 11 stations in the Old World that do likewise from Haparanda, Sweden, (65.8° N) to Cape Town, Union of South Africa, (33.9° S). The times of epoch (low) in barometric pressure for these stations with respect to the 2 wave lengths yield 26 timings each.

The 3.6296 Quarter Wave and Latitudinal Passage

Epochs (lows) with respect to the 3.6296 quarter wave length are plotted on the epoch chart in FIGURE 10. Entered at 20° latitude as before is the timing for this same wave length in sunspots. Added is a passage line, also as before. Here again we have an epoch chart on which have been plotted by angular measure epochs from the solar and terrestrial spheres. The passage line is a presumed one for the sun, but the lows for the same wave length in barometric pressure align themselves around this passage line from the sunspots as though related to it in some way. The poleward epochs tend to be the early ones, the equatorward epochs the late ones.

I obtained geomagnetic latitudes for the 26 stations of FIGURE 10 and plotted a second epoch chart using these geomagnetic latitudes (FIGURE 11). The wing-shaped design varies but little from the one that uses geographic latitudes for these same epochs. In FIGURE 10, 18 of the 26 fall in the half of the wave length centered upon the presumed passage line. This changes but little in FIGURE 11, where 19 of the 26 fall within the centered half of the wave length, not enough difference to indicate anything one way or the other.

The 4.4635 Quarter Wave and Latitudinal Passage

FIGURE 12 uses the same series of data rearranged with respect to the 4.4635 quarter wave length. The same transfer by angular measure has been done for the timing of high in sunspots with respect to this wave length. The timings of barometric pressure epochs (lows) with respect to this wave length occur earliest farthest poleward and latest farthest equatorward. They tend

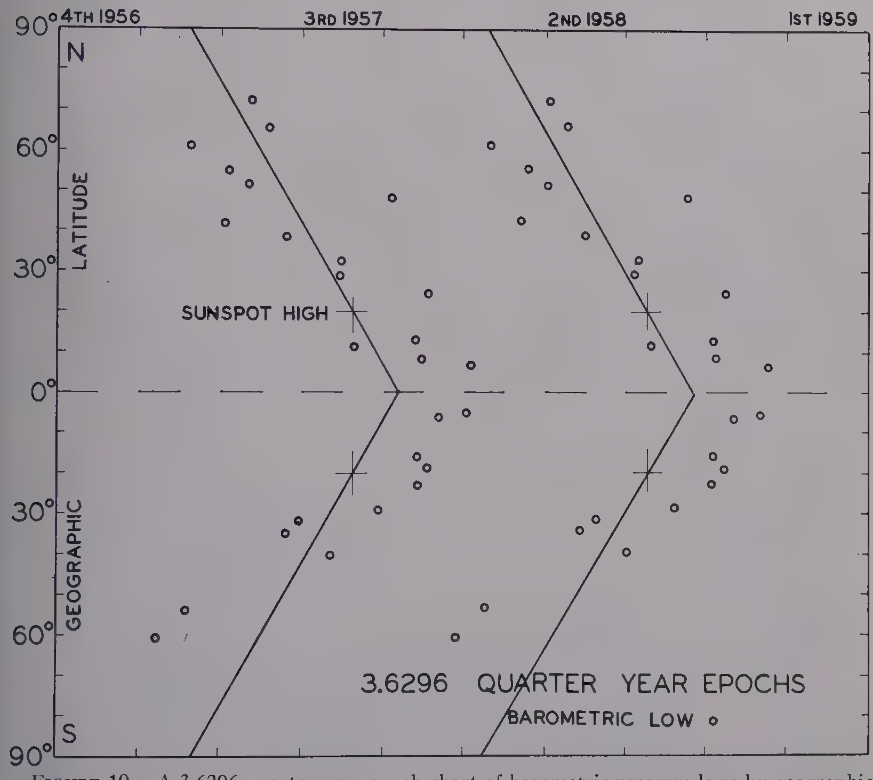


FIGURE 10. A 3.6296 quarter-year epoch chart of barometric pressure lows by geographic latitude. The passage line is computed from solar data and transferred bodily from the solar to the terrestrial sphere. The epochs of barometric pressure tend to cluster about the passage line.

to align themselves rather well around the presumed passage line of this wave length in the sunspots, which suggests an association of some sort.

FIGURE 13 presents by geomagnetic latitude the same epochs as in FIGURE 12. The wing-shaped diagram remains about as before, though a little less well coordinated. The pattern by geographic latitudes has 20 of the 26 epochs falling within the half of the wave length centered upon the presumed passage line (FIGURE 12). The pattern by geomagnetic latitudes (FIGURE 13), on the other hand, has 15 of the 26 falling within the centered half. The difference

appears enough to indicate a probable association greater by geographic than by geomagnetic latitudes.

The 5.115-Day Wave Length

A still shorter wave length, likewise of an extrasolar identification, is measured as 5.115-day wave length (Wing, 1959). This relatively short wave length has many advantages over longer ones. A full year of record, for

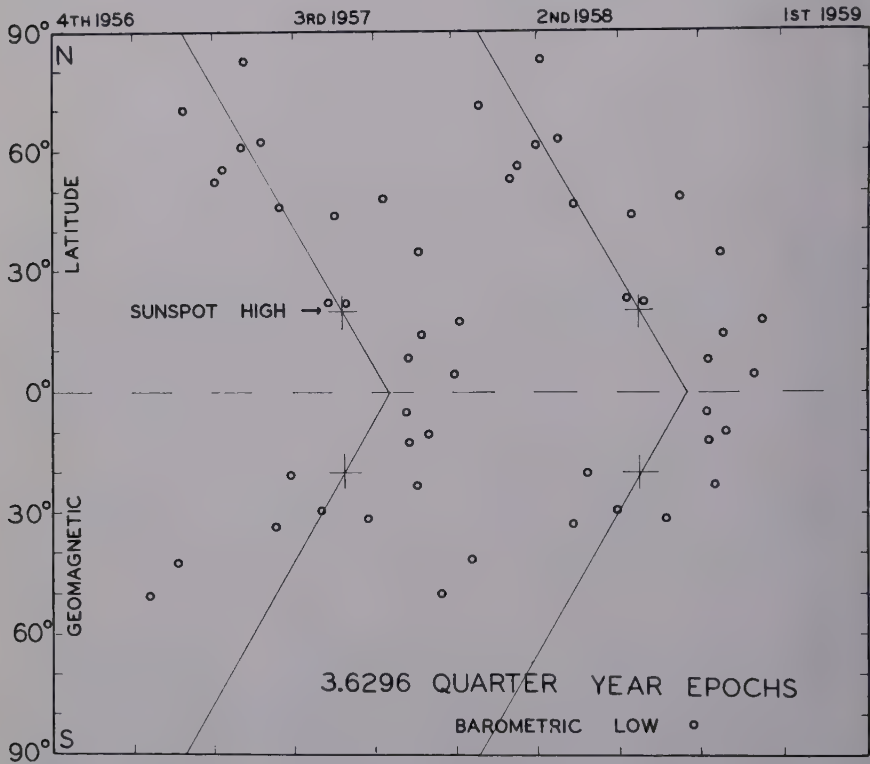


FIGURE 11. A 3.6296 quarter-year epoch chart of barometric pressure lows by geomagnetic latitude. The association is greater with geographic latitude as in FIGURE 10.

example, alone will give opportunity for more than 70 repetitions. It would take from scores of years to centuries for the same relative length of record for longer cycles. Long wave lengths studied in short records seem to involve a lot of guess work. Another important advantage is the wide latitude represented by the wonderful collections of the International Geophysical Year (IGY), nearly pole to pole, which can be used for a short wave length.

The 5.115-Day Wave Length and Barometric Pressure

I obtained at the IGY Center, National Weather Records Center, Asheville, N. C., noon barometric pressure readings for 39 representative stations

from Alert, N.W.T., Canada (82.5° N), to the South Pole (90.0° S). They form 2 transects by 10° zones of latitude, one roughly along the 75th meridian in the New World and the other along the 15th meridian in the Old World. But some variation had to be made, as in the inclusion of Australian stations in order to go south of Africa (TABLE 2).

Epochs (highs) with respect to this wave length are plotted on an epoch chart for calendar January 1959 (FIGURE 14). As before, a presumed passage

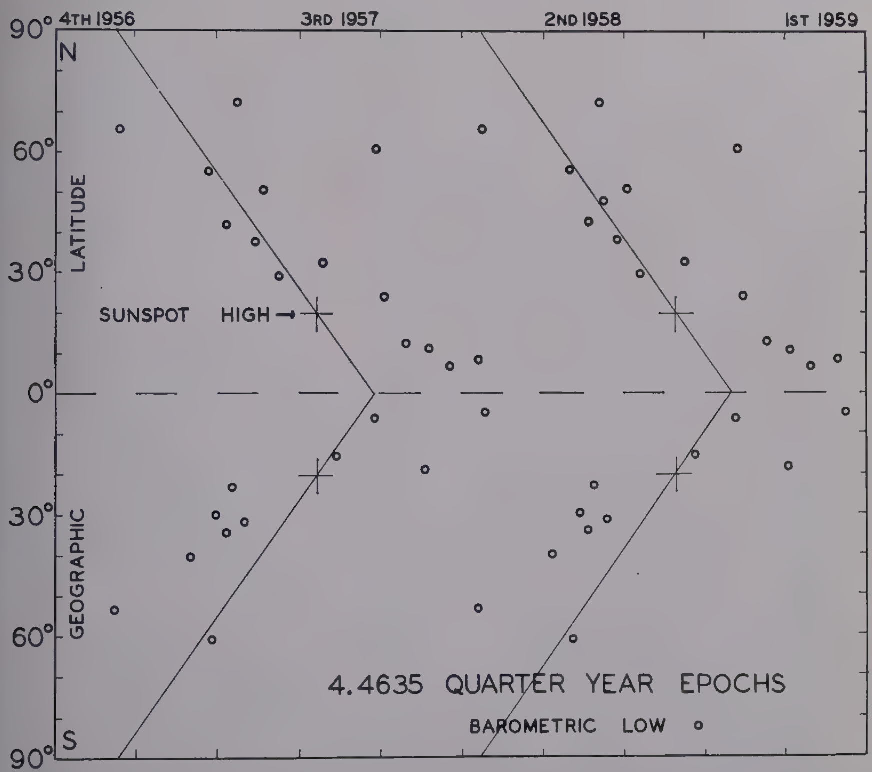


FIGURE 12. A 4.4635 quarter-year epoch chart of barometric pressure lows by geographic latitude. The passage line is computed from solar data and transferred bodily to the terrestrial sphere. The epochs of barometric pressure tend to cluster about the passage line.

line, with the ratio, also as before, for this wave length in the sunspots (timing of January 9.238, 1959) is added to the epoch chart. The highs in pressure with respect to the 5.115-day wave length form a band that tends to be earlier in the polar regions and progressively later equatorward. The band of epochs, which is terrestrial, tends to cluster at or near the presumed passage line of the same wave length in sunspots, which are solar.

FIGURE 15 is the same as FIGURE 14 except that it uses geomagnetic latitudes for one coordinate. This figure makes a looser form of band. For geographic latitudes, 25 of the 39 timings fall within the half of the wave length centered upon the passage line, while 14 fall in the remaining half. When geomagnetic

latitudes are used, however, 22 fall in the centered half of the wave length with 17 falling outside. At random, as many should fall in one half as in the other. However, association as measured by the difference between the 2 halves would appear more with geographic latitudes than with geomagnetic latitudes.

The 5.115-Day Wave Length and H-Magnetism

At the IGY Data Center in the Coast and Geodetic Survey, Washington, D.C., I was able to obtain 24 scaled, daily records of the horizontal compo-

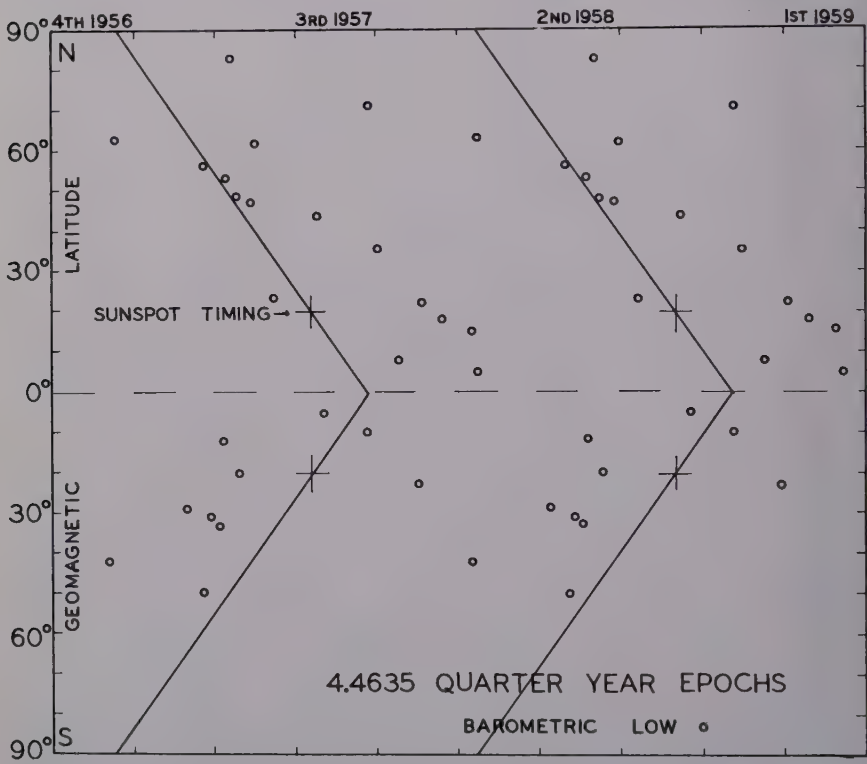


FIGURE 13. A 4.4635 quarter-year epoch chart of barometric pressure lows by geomagnetic latitude. The association is greater with geographic latitude as in FIGURE 12.

nent of terrestrial magnetism (H) ranging from Point Barrow, Alaska (71.3° N), to Marie Byrd Land, Antarctica, (80.0°). Each station record contains a full year or nearly so of data (TABLE 3).

I computed the timings of epoch (low) for the 24 stations and plotted them on an epoch chart (FIGURE 16). Again there appears a general pattern of earliest farthest poleward, progressively later equatorward. The presumed passage line in sunspots with respect to the 5.115-day wave length (as in FIGURES 14 and 15) has been placed upon this chart (FIGURE 16). The epochs tend to cluster around this passage line as though some way related, just as in the preceding figures.

However, the association with the passage line improves when plotted by geomagnetic latitude (FIGURE 17). Twenty of the lows fall in the half of the wave length centered upon the passage line, while 4 fall in the half not so

TABLE 2
EPOCHS WITH RESPECT TO 5.115-DAY WAVE LENGTH, BAROMETRIC PRESSURE

Station	Days in record	Geographic latitude °	Geomagnetic latitude °	Geographic longitude °	Time of epoch in month of Jan. 1959
Alert, B.C., Canada	363	82.5 N	85.8 N	62.3 W	6.005
Clyde, N.W.T., Canada	365	70.4 N	81.9 N	68.5 W	5.272
Resolution Island, New Zealand	362	60.3 N	71.3 N	64.9 W	9.562
Seven Islands, Quebec, Canada	365	50.2 N	61.2 N	66.3 W	8.632
Yarmouth, Nova Scotia, Canada	365	43.8 N	53.8 N	66.1 W	10.870
Bermuda	369	32.5 N	43.8 N	64.7 W	11.096*
Bonefish Bay, San Salvador	365	24.1 N	35.1 N	74.5 W	10.041
San Juan, Puerto Rico	365	18.4 N	29 N	66.0 W	8.738
Cuidad Bolivar, Venezuela	365	8.2 N	19.1 N	63.6 W	8.628
Ascension Island	359	7.9 S	1.4 S	14.4 W	9.513
Lima, Peru	356	12.0 S	1.0 S	77.0 W	7.655
La Quiaca, Peru	365	22.1 S	11.1 S	65.6 W	9.163
Cordoba, Argentina	365	31.3 S	20.3 S	64.2 W	8.713
Carmende Patagones, Argentina	365	40.8 S	29.8 S	63.0 W	7.945
Faro Cabo Blanco, Argentina	365	47.2 S	36.2 S	65.8 W	7.730
Ushuaia, Argentina	362	54.8 S	43.3 S	68.3 W	6.680
Destacamente, Argentina	364	64.3 S	53.3 S	63.0 W	6.705
Ellsworth, Antarctica	365	77.7 S	67 S	41.4 W	7.670
Byrd Station, Antarctica	365	80.0 S	70.6 S	120.0 W	4.247
Amundsen-Scott, Antarctica	375	90.0 S	78.5 S	—	6.055
Isfjord Radio, Norway	375	78.1 N	74.4 N	13.7 E	6.979
Andenes, Norway	365	69.3 N	67.1 N	16.1 E	5.247
Karlstad, Sweden	365	59.4 N	58.7 N	13.4 E	8.513
Praha-Ruzyne, Czechoslovakia	365	50.1 N	50.2 N	14.3 E	8.513
Foggia, Italy	365	41.4 N	41.3 N	15.5 E	8.987
Pantelleria, Italy	367	36.8 N	37.4 N	12.0 E	9.387
Hon, Libya	365	29.1 N	30.0 N	16.0 E	8.887
Bilma, Nigeria	364	18.7 N	20.3 N	12.9 E	8.987
Bousso, French Equatorial Africa	360	10.5 N	11.2 N	16.7 E	8.937
Makoua, French Equatorial Africa	365	0.5 S	0.3 N	15.9 E	11.995
Porto Amboim, Angola	365	10.7 S	9.1 S	13.8 E	8.105
Tsumeb, South West Africa	357	19.2 S	17.6 S	17.8 E	7.580
O'Okiep, Union of South Africa	354	29.6 S	28.1 S	17.9 E	9.512
Wagga Wagga, N.S.W., Australia	364	35.1 S	44.1 S	147.4 E	6.380
Gough Island	365	40.3 S	33.3 S	9.9 W	7.405
Macquarie Islands	365	54.5 S	61.1 S	159.0 E	6.870
Wilkes Station, Antarctica	365	66.2 S	77.8 S	110.6 E	4.397
Mawson, Antarctica	365	67.6 S	73.1 S	62.9 E	4.272
McMurdo, Antarctica	365	77.8 S	78.6 S	166.6 E	8.687

* August 29, 1957 to Sept. 1, 1958.

centered. For geographic latitudes, on the other hand, 18 fall within the centered half as against 6 that fall in the remaining half (FIGURE 16). If these results are indicative, it appears that latitudinal passage as expressed by the horizontal component of terrestrial magnetism seems associated more with geomagnetic latitude than with geographic latitude. This does not seem to be the case in other terrestrial phenomena. Additionally, H-magnetism seems

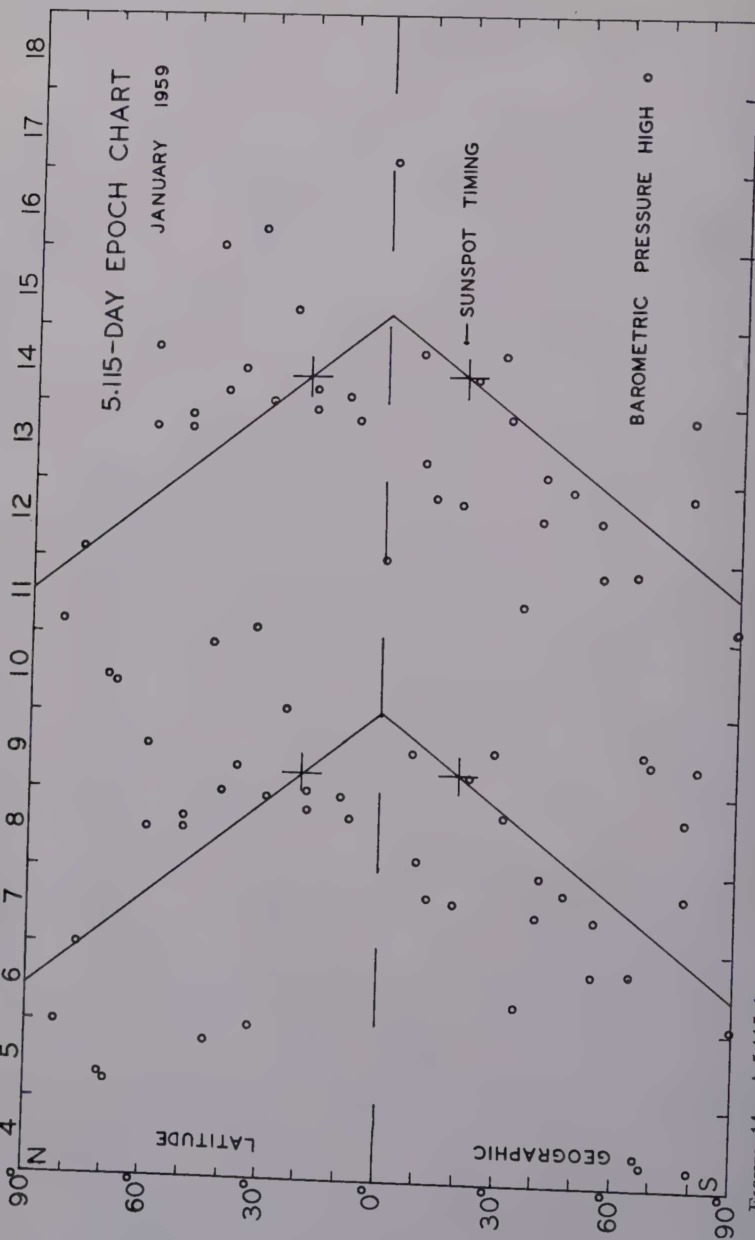


FIGURE 14. A 5.115-day epoch chart of barometric pressure highs by geographic latitude. The passage line is computed from solar data. The epochs of barometric pressure tend to cluster about the passage line.

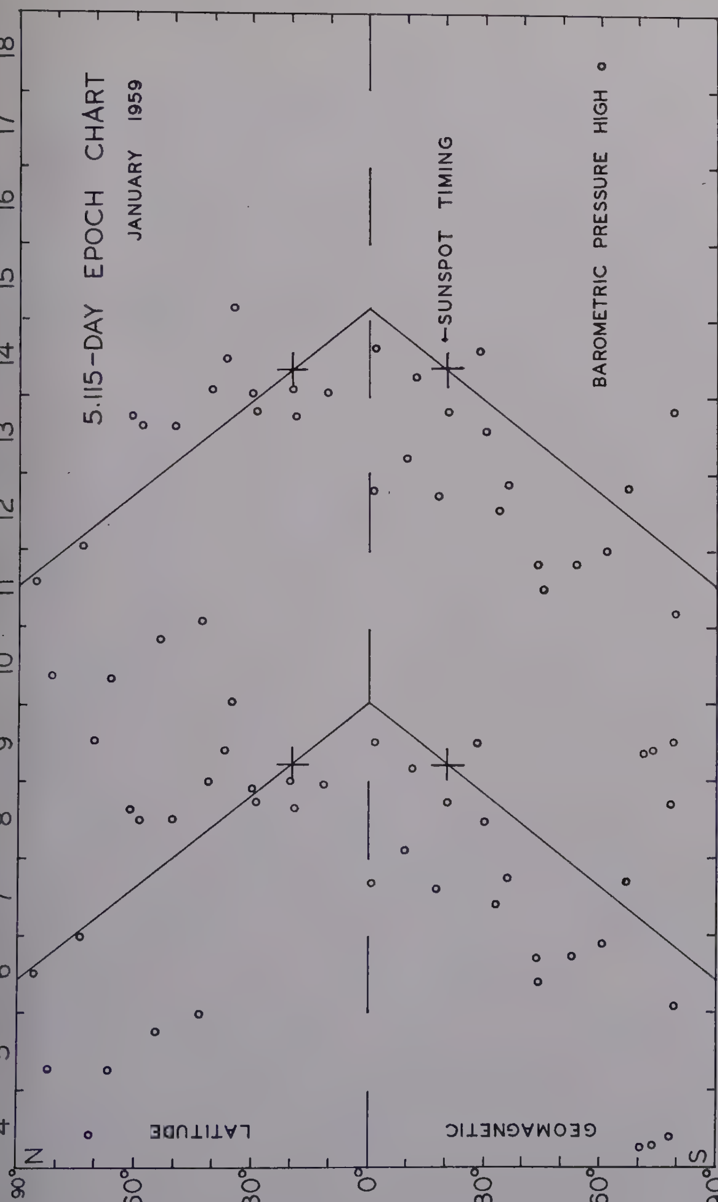


FIGURE 15. A 5.115-day epoch chart of barometric pressure highs by geomagnetic latitude. The passage line is as in FIGURE 14. The association is greater with geographic than with geomagnetic latitude.

somewhat more closely associated with the presumed sunspot passage line than does barometric pressure, a greater proportion of its epochs falling closer to the passage line.

TABLE 3

EPOCHS WITH RESPECT TO 5.115-DAY WAVE LENGTH, H-MAGNETISM

Station	Number of unqualified items	Geographic latitude	Geomagnetic latitude	Geographic longitude	Timing of low as of Greenwich time (Jan. 1959)	
Point Barrow, Alaska	354	71.3 N	68.6 N	156.8 W	8.257	
College, Alaska	365	64.8 N	64.7 N	147.8 W	6.497	
Scrednikan, U.S.S.R.	415	62.4 N	53.2 N	152.3 E	9.020	Oct. 1, 1957–Dec. 30, 1958
Lovo, Sweden	367	59.4 N	58.2 N	17.8 E	9.387	
Sitka, Alaska	364	57.0 N	60.0 N	135.3 W	8.280	
Eskdalemuir, Great Britain	364	55.3	58.4 N	3.2 W	8.858	
Vienna, Austria	365	48.3 N	47.9 N	16.3 E	9.291	
Beloit, Kansas	275	39.4 N	49.3 N	98.1 W	8.547	
Kakioka, Japan	367	36.2 N	26.0 N	140.2 E	8.891	
Tucson, Arizona	365	32.2 N	40.4 N	111.0 W	8.814	
Honolulu, Hawaii	365	21.3 N	21.0 N	158.1 W	9.670	Dec. 1, 1957–Nov. 30, 1958
Bombay, India	365	18.6 N	9.5 N	72.9 E	10.904	July 1, 1957–June 30, 1958
San Juan, Puerto Rico	365	18.4 N	29.9 N	66.0 W	8.197	
Trivandrum, India	347	8.5 N	1.1 S	77.0 E	8.779	
Fuguene, Columbia	362	5.5 N	16.9 N	73.8 W	12.228	July 1, 1958–July 31, 1959
Talera, Peru	357	4.6 S	6.6 S	81.3 W	11.484	
Pulan Burung, Indonesia	362	6.0 S	12.1 S	106.8 E	9.692	July 1, 1957–June 30, 1958
Apia, Samoa	357	13.8 S	16.0 S	171.8 W	8.937	Nov. 1, 1957–Oct. 31, 1958
Hermanus, Union of South Africa	367	34.4 S	33.3 S	19.2 E	8.987	
Trelew, Argentina	364	43.2 S	31.7 S	65.3 W	9.437	
Wilkes Station, Antarctica	363	66.2 S	77.8 S	110.6 E	6.280	
Halley Bay, Antarctica	370	75.0 S	65.8 S	26.6 W	8.712	May 1, 1957–May 7, 1958
Little America, Antarctica	359	78.2 S	74.0 S	162.2 W	6.780	
Marie Byrd Land, Antarctica	268	80.0 S	70.6 S	120.0 W	5.797	Aug. 23, 1957–June 30, 1958

Passage and Counter Passage

It will be recalled that some of the solar prominences, those of higher solar latitudes, have an apparent passage behavior from lower to higher latitudes, backwards to that in the sunspots (FIGURE 3). When plotted on an epoch chart, they show behavior seemingly reciprocal to that of the sunspots themselves. In FIGURE 3 the poleward passage line is reciprocal to the presumed equatorward passage line of the sunspots for 90° of latitude plotted with the timing of 1948.68 at 70° (70° is the position reciprocal to 20°). That the reciprocal line fits the solar prominences rather well seems obvious. It seems

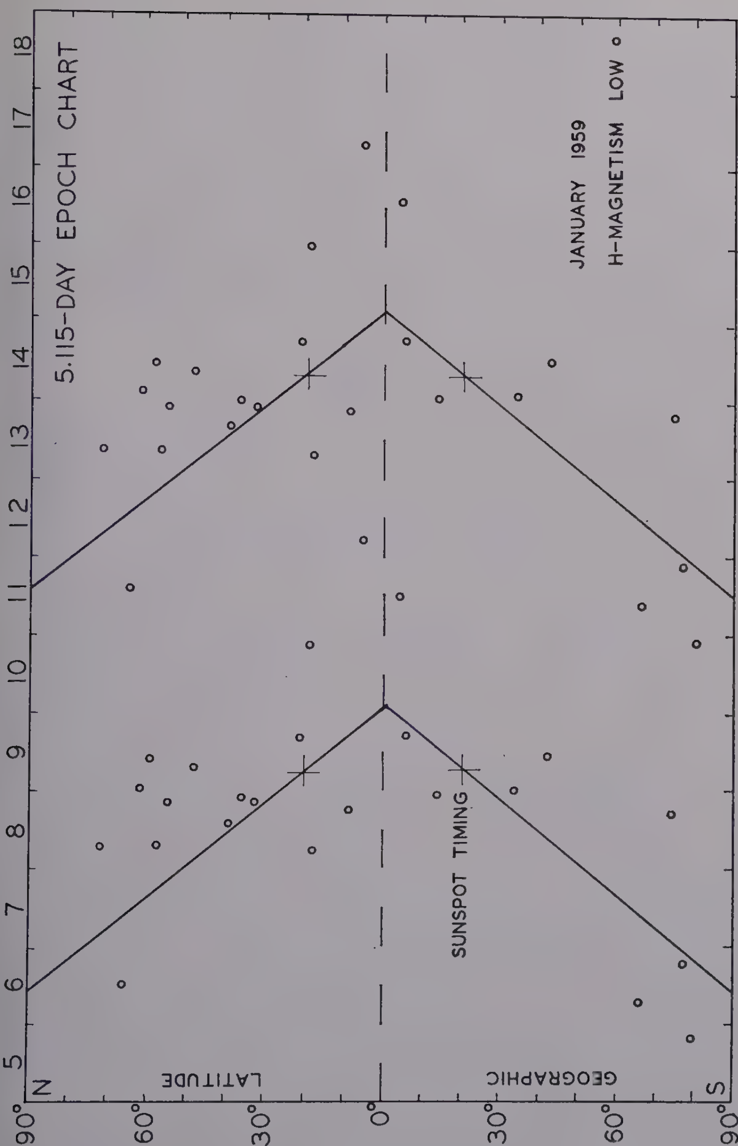


FIGURE 16. A 5.115-day epoch chart of H-magnetism lows by geographic latitude. The passage line is as in FIGURE 14.

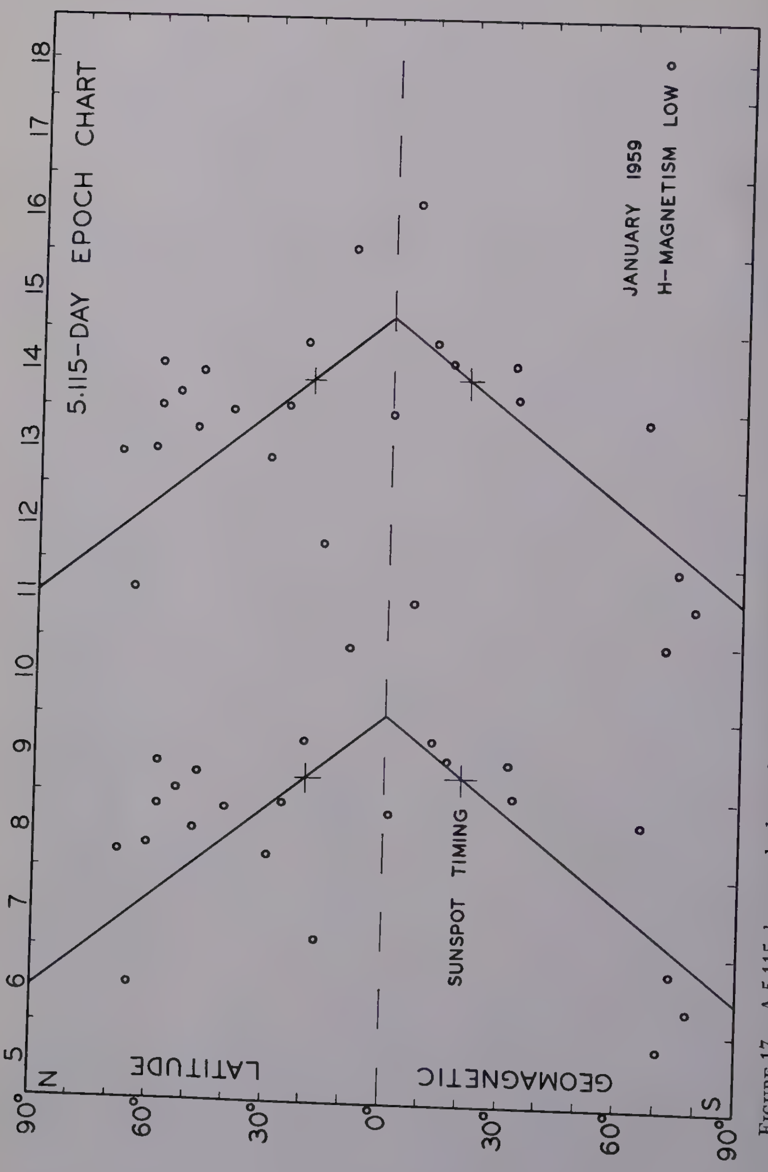


FIGURE 17. A 5.115-day epoch chart of H-magnetism lows by geomagnetic latitude. The passage line is as in FIGURE 14. H-magnetism is the only phenomenon yet tested whose epochs are associated more with geomagnetic than with geographic latitudes.

to indicate a reciprocal relationship of some kind, but what that may be is not at all clear.

As mentioned earlier in this paper, because passage on earth for all cycles tested thus far runs from the poles to the equator, seemingly simultaneously with similar behavior on the sun, it really does not take a great deal of scientific imagination to conceive also of a possible reciprocal passage on earth for each wave length. Solar prominences sometimes rise to great heights, thousands of miles above the solar surface. The reciprocal passage on the sun with respect to the 11.08-year wave length, therefore, seems to involve phenomena of high solar altitudes. An intriguing idea rises: the possibility of similar behavior at high terrestrial altitudes. This idea surely merits testing, but what can be examined and how the testing should be done is a problem. Again, we are indebted to the International Geophysical Year for data that make possible one such test.

The 5.115-Day Wave Length and the F-Layer

At the IGY Data Center in the National Bureau of Standards Laboratory, Boulder, Colo., I obtained noon values for the critical frequency of the F2 layer (foF2) for 38 stations ranging from Fletcher's Ice Island, Arctic Ocean (about 79° N) to the Amundsen-Scott station at the South Pole (90° S). Four heights of the F-layer (hF) supplement the 38 of the foF2. All these represent altitudes presumably above 100 to 150 miles. Very great difficulties evidently faced the collectors in obtaining the readings because of the hornets' nest of disturbances bearing upon the readings. This is particularly true of stations in higher latitudes, where the station operators qualified a great many items. I therefore used only such data as the operators themselves did not question.

If there is any kind of counter passage associated with the earth like that suggested on the sun by solar prominences, it might be expected to appear perhaps in the ionosphere. For the 5.115-day wave length, the pattern of distribution of epochs should be opposite to that in barometric pressure and H-magnetism at the surface (see FIGURES 14 to 17). Epochs should align themselves about the reciprocal passage line. (By reciprocal, I mean a passage line with the same timing as before but posted as at 70° latitude instead of 20° latitude, the counter passage line running poleward through this point from the equator to the poles. The previous timing at the poles will now be that of the equator and *vice versa*.)

Accordingly, I determined the times of epoch (high) with respect to the 5.115-day wave length in the ionosphere data. I have plotted them on the same base as before (FIGURE 18). I have added a reciprocal passage line based upon timing in the sunspots (that is, upon January 9.238, 1959 at 20° latitude). The reciprocal passage line goes from the equator to the poles, passing through this timing of solar high but now posted at 70° latitude. The passage line and timing has been transferred bodily, in a sense, from the solar to the terrestrial sphere. I have entered upon the epoch chart two previously determined timings (indicated by crosses), those of Washington and Huancayo for 1950 (Wing, 1957) brought down to early January 1959 in multiples of

5.115 days. Their falling as they do within the current pattern would be expected if the 5.115-day wave length is a continuous one. The pattern of distribution of the forty epochs and their tendency to associate with the reciprocal passage line seems rather clear. In so far as this wave length is concerned at least, counter passage in the ionosphere would seem demonstrated. It suggests by analogy a similar counter passage behavior with respect to this

Table 4
Epochs with Respect to 5.115-Day Wave Length, F-Layer

Station	Number of unqualified items	Geographic latitude	Geo-magnetic latitude	Time zone	Timing of high at site (Jan. 1959)	Timing of high, Greenwich time (Jan. 1959)
Fletcher's Ice Island*	94	79 N	84 N	120 W	8.080	8.413
Fletcher's Ice Island	27	79 N	84 N	120 W	8.838	9.171
Thule, Greenland*	176	76.6 N	88 N	75 W	8.938	9.146
Thule, Greenland	123	76.7 N	88 N	75 W	7.805	8.013
Point Barrow, Alaska	146	71.3 N	68 N	150 W	8.963	9.380
Fairbanks, Alaska	196	64.9 N	64 N	150 W	9.188	9.605
Adak, Alaska	203	51.9 N	48 N	180 W	7.480	7.980
St. Johns, Nfld., Canada	242	47.6 N	58.4 N	60 W	9.288	9.455
San Francisco, Calif.	176	37.4 N	43 N	120 W	6.780	7.113
White Sands, N.M.	237	32.3 N	41 N	105 W	7.830	8.122
Grand Bahama, Bahama Islands	217	26.6 N	38 N	75 W	7.330	7.538
Maui, Hawaii	241	20.8 N	21 N	150 W	5.660	6.077
Puerto Rico	239	18.5 N	29 N	60 W	7.330	7.497
Panama Canal Zone	282	9.4 N	20 N	75 W	6.105	6.313
Chiclayo, Peru	109	6.8 S	4.4 N	75 W	7.960	8.168
Huancayo, Peru	133	12.0 S	1 S	75 W	7.748	7.956
Tucuman, Argentina	137	26.9 S	16 S	60 W	6.130	6.297
Buenos Aires, Argentina	326	34.5 S	23 S	60 W	7.830	7.997
Port Stanley, Falkland Is.	192	51.7 S	41 S	60 W	9.288	9.455
Wilkes Station, Antarctica*	189	66.2 S	78 S	105 E	7.838	7.546
Wilkes Station, Antarctica	55	66.2 S	78 S	105 E	9.618	9.326
Ellsworth, Antarctica	51	77.7 S	67 S	45 W	9.863	9.988
South Pole*	187	90.0 S	78.5 S	0	9.838	9.838
South Pole	88	90.0 S	78.5 S	0	10.084	10.084
Longyearbyen, Spitzenbergen	177	78.2 N	74 N	15 E	11.820	11.778
Lychsele, Sweden	267	64.6 N	63 N	15 E	9.080	9.038
Moscow, U.S.S.R.	271	55.5 N	51 N	30 E	10.670	10.587
Wakkanai, Japan	162	45.2 N	35 N	135 E	7.855	7.480
Kokubungi, Japan	175	35.7 N	26 N	135 E	7.938	7.563
Okinawa	265	26.3 N	15 N	135 E	8.905	8.530
Baguio, Philippine Islands	231	16.4 N	5 N	120 E	7.430	7.097
Kodaikanal, India	186	10.2 N	0.7 N	75 E	5.730	5.522
Singapore	153	1.3 N	10 S	105 E	6.180	5.888
Leopoldville, Congo	184	4.3 S	3.1 S	15 E	6.405	6.363
Townsville, Australia	151	19.3 S	28 S	150 E	7.080	6.663
Brisbane, Australia	207	27.5 S	36 S	150 E	8.730	8.313
Canberra, Australia	245	35.3 S	44 S	150 E	6.880	6.463
Godley Head	280	43.5 S	49 S	180 E	10.463	9.963
Campbell Island	143	52.5 S	57 S	165 E	10.870	10.412
Macquarie Island	81	54.5	60 S	150 E	10.113	9.696
Cape Hallett, Antarctica	67	72.3 S	76 S	165 E	8.030	7.572
Scott Base, Antarctica	161	77.8 S	79 S	165 E	9.963	9.505

* Height of F-layer; other entries are for critical frequency of F-2-layer.

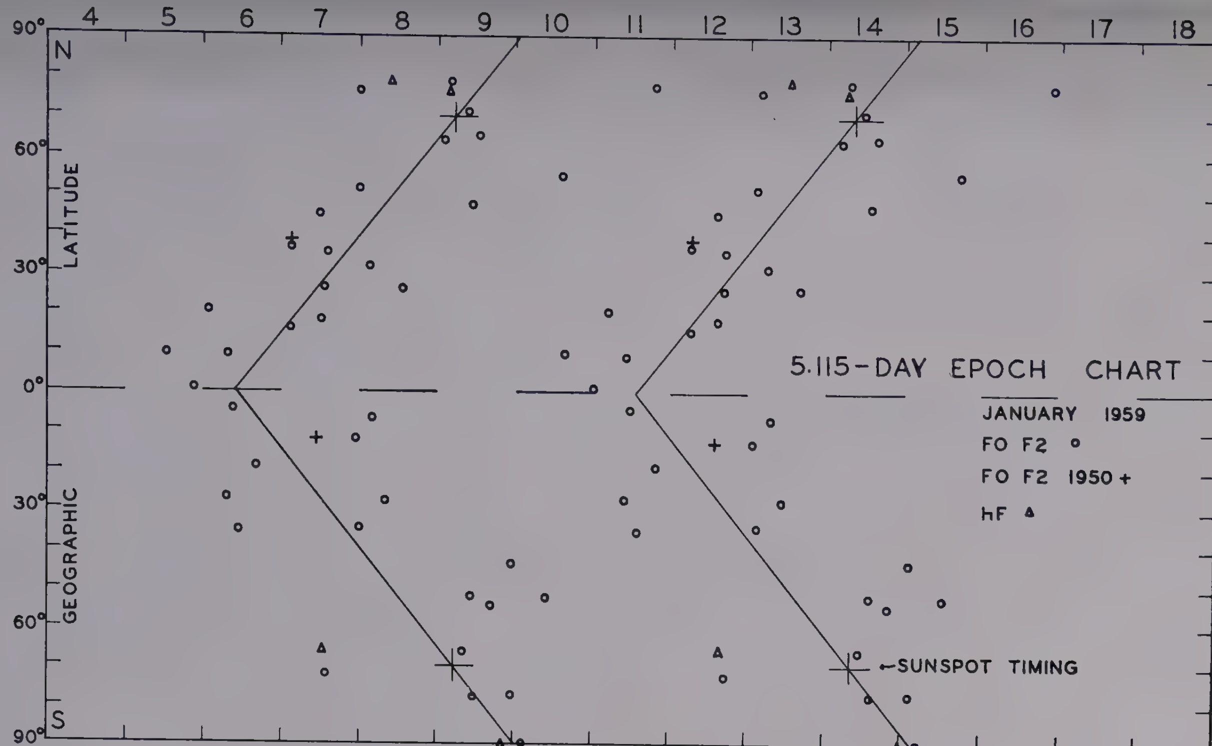


FIGURE 18. A 5.115-day epoch chart of ionosphere timings of high by geographic latitude. The passage line is reciprocal to that of FIGURE 14. The ionosphere epochs seem associated with this reciprocal passage line. It suggests a behavior at high terrestrial altitudes parallel for its wave length with thep oleward passage in solar prominences in FIGURE 3. The wing-shaped design is backwards in the ionosphere from that at the surface. The earlier epochs are farther equatorward, the later ones farther poleward.

wave length at high solar altitudes also, presuming that synchronous behavior occurs with angular relationships over the solar and terrestrial spheres. It must be remembered, of course, that the 5.115-day wave length is but a run-of-the-mill one; it does not seem particularly prominent or important in either solar or terrestrial affairs. How it and others combine their strength and weaknesses at various times and in various latitudes, however, may be another matter.

That it parallels for its wave length, but over almost a full 90° of latitude in each hemisphere, the behavior of solar prominences with respect to the 11.08-year wave length seems to suggest the possibility of counter passage for *any* wave length having equatorward passage. Presumably, this would mean any and all wave lengths.

This offers a possible explanation also for the association of aurora highs in terrestrial polar regions with the 11-year manifest cycle in the sunspots, themselves of low solar latitudes. The association may be with a counter passage phenomenon of the 11-year manifest cycle rather than directly with the sunspots themselves. It would seem possible to test this matter if past auroras can be pinpointed by latitude in sufficient numbers. We face the circumstances, however, that such reports may reflect the latitudes of the observers themselves more than the latitudes of the auroras.

The passage behavior of epochs of H-magnetism seem associated more with geomagnetic latitude than with geographic latitude. Temperature and barometric pressure epochs, on the other hand, seem associated more with geographic latitude than with geomagnetic latitude. Which if either may be the association of the ionosphere epochs (that is, F-layer highs)?

I prepared an epoch chart of the ionosphere epochs (F-layer highs) with respect to the 5.115-day wave length using the respective geomagnetic latitudes of the several stations (FIGURE 19). While the pattern remains generally the same as that for geographic latitudes in FIGURE 18, the association with the reciprocal passage line is less. By geographic latitude, 32 of the 42 timings fall within the half of the wave length centered upon the presumed reciprocal passage line; 10 fall in the half not so centered (omitting the two 1950 epochs now in current time). By geomagnetic latitude, on the other hand, 29 fall within the centered half of the wave length; 13 fall in the half not so centered. It may be, of course, that some revision would occur if some latitudes of reference other than 20° and 70° were used. Additionally, 20° and 70° geographic latitudes may have some other geomagnetic representation than 20° and 70° geomagnetic latitude. However, I do not see how any great change in the pattern would be made.

Geomagnetic Versus Geographic Latitudes

It would seem therefore that passage in H-magnetism epochs is related to geomagnetic latitude, but that such passage in temperature, barometric pressure, and ionosphere epochs is not. The association of the epochs with geographic and geomagnetic latitudes or the lack of association with one or the other puzzles me, which may indicate only how ignorant I may be. Other than for H-magnetism, epochs of terrestrial phenomena thus far tested seem

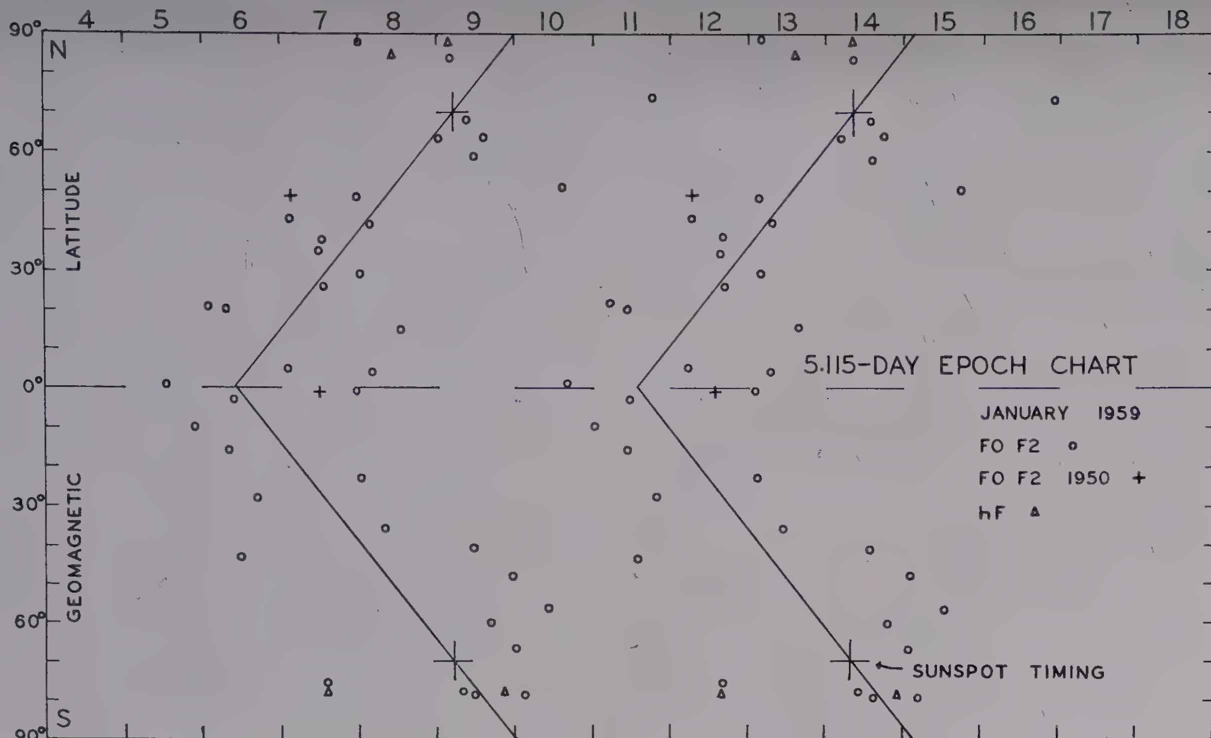


FIGURE 19. A 5.115-day epoch chart of ionosphere timings of high by geomagnetic latitude. The passage line is as in FIGURE 18. The epochs show a greater association with geographic than with geomagnetic latitude.

associated more with geographic than with geomagnetic latitude. This being the case, it would appear that the mechanism through which the cycles are working hardly operates wholly through the magnetic field as identified with geomagnetic latitudes. At least I do not find any studies indicating that the operations of the magnetic field are associated more with geographic than with geomagnetic latitude.

There is one factor that may alter the picture. The epochs that I have used in FIGURES 10 to 19 are of stations that form rough latitudinal transects. Perhaps stations from all longitudes might change the picture of greater association with geographic than with geomagnetic latitude. Thus what appears to be a behavior with respect to geographic latitude may not be so were I to use stations in more different longitudes. However FIGURES 8 and 9 (4.222-year wave length in temperature) argue against this idea, as they include nearly all longitudes. This is certainly a matter in need of more study.

Generalized Passage and Counter Passage

I think that it is possible to generalize on the behavior described as latitudinal passage and counter passage. FIGURE 20 represents a sphere that could be either the sun or the earth. Perhaps it might just as well be any other star or planet. Arrows indicate the equatorward passage at the surface and counter passage at high altitudes, passage and counter passage being at the presumably common rate of 90° passage = $\sqrt{\tfrac{1}{2}\,(\text{wave length})^2}$. On earth, somewhere between the height of temperature, barometric pressure, and magnetic readings (seldom more than about 6 feet above the surface) and the height of the F-layer, equatorward passage gives way to poleward passage, perhaps in some sort of belt of neutrality. Perhaps its height may vary with different wave lengths. Of this we know nothing. Somewhere on the far side of the F-layer, counter passage would presumably fade out into nothingness. Of this we know nothing either. Obviously, it is important to find out if these things are true and where they may happen.

On earth, passage may be traced for full 90°, poles to equator in both hemispheres (at least 82.5° N); on the sun, passage seems traceable only in low middle latitudes, roughly from 40° to the solar equator. On earth, counter passage seems traceable for full 90°, equator to poles in both hemispheres (at least 79° N); on the sun, counter passage seems traceable only from low-middle latitudes to the polar regions. On the earth, passage and counter passage may be traced for many wave lengths, perhaps eventually for any and all. On the sun, the extraordinarily dominant 11.08-year wave length for some pygmatic reasons overwhelms everything else. On the sun, passage and counter passage are invisible or inoperative in the respective zones where we see no sunspots or no prominences. If analogy with terrestrial behavior is any guide, presumably the phenomenon itself continues in these zones of invisibility, whether or not we see it expressed.

It seems possible to theorize therefore that cyclic controls in some way pass like ripples over the solar and terrestrial spheres, the passage rates being proportional to the wave lengths. How many such cycles there may be, one can hardly guess. Many surely do exist, perhaps a continuous series of wave

lengths from extremely short ones to extremely long ones. I know of no physical mechanism to account for this behavior.

The three wave lengths of extrasolar identification have been measured in the light curves of three different variable stars (Wing, 1959). They have been arbitrarily used as some wave lengths for testing solar and terrestrial data. Their apparent presence suggests several possibilities. For one thing,

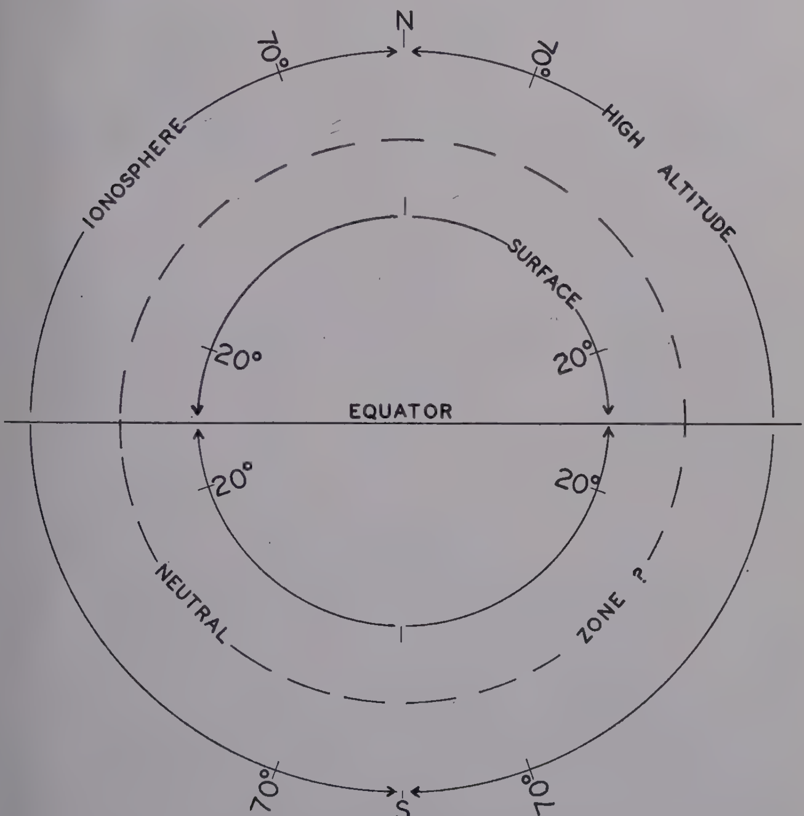


FIGURE 20. Generalized concept of latitudinal passage and counter passage for a sphere, either the solar or terrestrial. Counter passage moves poleward at high altitudes, passage moves equatorward at the surface. Somewhere between would seem to be a neutral zone. Beyond the ionosphere, for example, counter passage would fade out.

the wave lengths and their terrestrial association may be wholly accidental, fortuitous behavior, but this would seem to stretch credulity beyond reasonable limits. They may reflect jigglings or vibrations in the solar system or in the universe itself. They may be independent things, like the wave lengths of light from a match struck here on earth identical with those of a distant star. A last possibility is that they may represent wave lengths that in their history have touched respectively upon distant, variable stars and the solar-terrestrial systems. Their passing may be duly noted by ups and downs in a

host of solar-terrestrial phenomena. The sunspots, rainfall, pressure, and tree rings may be but a few of the many events that register their passing. These things then may be the Geiger counters, the gauges, and the instruments that record their passing.

The phenomena described in this paper seem to demonstrate latitudinal passage (and latitudinal counter passage) for several wave lengths varying from 5.115 days to 11.08 years, whatever may be their generating source. Presumably there should be wave lengths measured in fractional days also, which would mean in hours. If in hours, we might well expect wave lengths measured in minutes and fractional minutes, which would mean in seconds. In the other direction, they are measurable in years; presumably, they should be measurable in decades and centuries also. What the upper and lower limits to expect may be, I have no way of even hazarding a guess. That they are longer than 11.08 years and shorter than 5.115 days, I feel sure.

Summary

This paper describes certain solar and terrestrial behavior that seems to indicate a general principle.

The equatorward shift of sunspots and poleward shift of some solar prominences with respect to the 11-year manifest cycle in sunspots appear to have a reciprocal relationship.

The equatorward shift is termed *latitudinal passage*; the poleward shift is termed *counter passage.*

The dominant wave length in sunspots is established as 11.08-years wave length.

The variable dominance of different wave lengths in different solar and terrestrial phenomena is described as pygmatics. A dominant wave length in one phenomenon may be a run-of-the-mill cycle in others. All wave lengths of cycles measured in terrestrial phenomena appear to be present also in solar events, particularly the sunspots.

Tests of terrestrial data of temperature, precipitation and runoff, barometric pressure, and H-magnetism variously with wave lengths of 11.08 years, 9.60 years, 4.222 years, 4.4635 quarter-years, 3.635 quarter-years, and 5.115 days indicate that latitudinal passage occurs on the earth, from pole to equator seemingly simultaneously with a similar latitudinal passage on the solar sphere.

The ratio of passage seems always the same for all wave lengths and is established as 90° passage = $\sqrt{\tfrac{1}{2}\,(\text{wave length})^2}$.

Thus the winged-shaped or butterfly pattern reported in sunspot behavior appears in terrestrial events also.

It suggests a behavior passing ripplelike or wavelike over the solar and terrestrial spheres, poles to equator.

Tests of the ionosphere (F-layer) indicate counter passage at high terrestrial altitudes, reciprocal to passage equatorward at the surface.

Terrestrial passage in H-magnetism seems associated with geomagnetic latitude. Passage in all other events tested seems associated more with geographic than with geomagnetic latitude.

The several tests indicate a general principle that epochs of the various

cycle wave lengths, whatever their generant may be, appear respectively first in the polar regions of the earth and sun and pass equatorward, passage being simultaneous wave length for wave length and latitude for latitude on the earth and sun.

A counter-passage behavior reciprocal to equatorward passage seems present at high solar and terrestrial altitudes.

Somewhere between the height of terrestrial thermometers, rain gauges, barometers, and magnetometers and the height of the F-layer should occur some sort of neutrality belt where equatorward surface passage gives over to poleward counter passage.

Many rises and falls in solar and terrestrial events, such as sunspots, temperature, and rainfall, and a host of others, seem to be recording the passing of waves, whatever they may be and from whatever source they may come.

Acknowledgments

For assistance in obtaining data my thanks are due to William McMurray, Richard Davis, Mary Browning, and others of the IGY Center, National Weather Records Center, Asheville, N.C.; to Robert A. Earle, J. W. Gettemy, William Paulishak, and others of the IGY Center in the Coast and Geodetic Survey, United States Department of Commerce, Washington, D.C.; and to Walter B. Chadwick, S. M. Ostrow, Patricia Pardi, and others of the IGY Center in the National Bureau of Standards Laboratory, Boulder, Colo. My thanks are due also to Y. M. Yerdjevich for data pertaining to the Göta, Nemunas, Rhine, and Danube rivers contained in TABLE 1.

References

CLAYTON, H. H. & F. L. CLAYTON. 1944 & 1947. World Weather Records. Smithsonian Misc. Collections. 79, 90, 105. Smithsonian Inst. Washington, D.C.

DEWEY, E. R. 1958. The length of the sunspot cycle. J. Cycle Research. **7:** 79–81.

DEWEY, E. R. 1960. The 200-year cycle in the length of the sunspot cycle. J. Cycle Research. **9:** 67–92.

JARVIS, C. S. 1953. Flood-stage records of the river Nile. J. Cycle Research. **2:** 96–100.

SCHOVE, D. J. 1956. Sunspot maxima since 649 B.C. J. Brit. Astronomical Assoc. **66:** 59–61.

SHOSTAKOVICH, W. B. 1934. Mud deposits in lakes and the periodic fluctuation phenomena in nature. Zapiski State Hydrologic Inst. **13:** 95–138.

WING, L. W. 1954. Global pattern of 4.222-year cycles in temperature. J. Cycle Research. **3:** 55–83.

WING, L. W. 1956. Ultra-long waves and global patterns of temperature. *Ibid.* **5:** 31–59.

WING, L. W. 1957. Ultra-long waves and solar-terrestrial cycle relationships. *Ibid.* **6:** 55–119.

WING, L. W. 1958. Latitudinal passage behavior. *Ibid.* **7:** 67–78.

WING, L. W. 1959. Solar-terrestrial relations indicating possible cosmic waves. *Ibid.* **8:** 67–95.

WING, L. W. 1960. Scattered reports and latitudinal passage in rabbit and rodent cycles. *Ibid.* **9:** 51–66.

INVESTIGATIONS OF MILANKOVITCH AND THE QUATERNAR CURVE OF EFFECTIVE SOLAR RADIATION

Wenceslas S. Jardetzky
Manhattan College, New York, N.Y.

Secular variations of solar radiation reaching the earth became a mathemati cal problem in a paper by Meech.[1] He made the first attempt to "throw ligh upon geological researches relating to changes of the heat of the globe at ver remote epochs," and developed formulas relating the radiation at any latitud and its secular variations to changes in eccentricity, longitude of the peri helion, and inclination of ecliptic. The results of Meech were rediscovere and completed by Wiener in 1877 and Hargreaves in 1896, but the mathemati cal analysis of the problem was greatly advanced by the investigations (191 to 1941) of Milankovitch.[2-5]

The first important work of Milankovitch was published in 1920. Its ai was to present a mathematical theory of a number of thermal phenomen produced on the earth by solar radiation. The starting point was a simpl physical problem that yields the amount of solar radiation reaching a uni area of the earth's surface. The complexity of those thermal phenomen makes it necessary to investigate separately the part played by many factor In his investigation of 1920, Milankovitch developed a restricted mathemati cal theory of the climate of the planet leading him to the concept of the sola climate, that is, the climate that depends exclusively on the amount of sola radiation reaching a place according to its geographical latitude. Severa assumptions made by Milankovitch show in a precise manner the restriction used to make a mathematical treatment of the problem possible. It is evi dent that the basic assumptions are related to two groups of phenomena astronomical and geophysical, as follows.

(1) The intensity of solar radiation could vary during the geological period Therefore an assumption must be made about the solar constant. Meech, fo example, assumed that the physical constitution of the sun remained constan for the time interval of 100,000 years, which he considered.

(2) The most essential changes in the earth's position with respect to th sun and, therefore, in radiation during the last 650,000 years, are due chiefly t variations of three elements of the earth's orbit: the longitude of the peri helion, eccentricity, and inclination of the ecliptic. The fact that the radia tion depends essentially on the combined variations of these elements was no always understood well enough in the past. Moreover it is necessary to mak an assumption about the time interval for which the corresponding astronom cal data are reliable. It seems that celestial mechanics yields adequate dat for about the last 1 million years.

(3) With the help of these data, the solar radiation and its variations fo any geographical latitude can be expressed in terms of three astronomical el ments and their variations. The intensity of radiation can be given in gm cal. or in special units (for example, the "thermal days" of Meech). "Canon cal units," used by Milankovitch, were defined by assuming the solar consta

($J_0 = 2$ gm. cal./cm.^{2} min.) to be equal to unity and by introducing a time unit equal to $\frac{1}{100,000}$ of the tropical year.

Among the applications at the first stage of the theory developed by Milankovitch, we find the calculation of the mean temperature of the lower layer of the earth's atmosphere (15.2° C., differing by 0.1° C. from the observed value). Similar calculations were made for the temperatures of Mars and Mercury. Also very important was a relationship between a variable radiation and the temperature of the ground. At the suggestion of W. Köppen, Milankovitch began in 1922 the computation of secular variations of the effective solar radiation for three latitudes: 55°, 60°, 65° N for the last 650,000 years. Köppen was the first to see a possible connection between the radiation diagram given by Milankovitch and an explanation of the ice ages. The curves representing the variations of radiation can be given in two forms: either by plotting the intensity of radiation against the time, or by using equivalent latitudes for every year. For example if we consider the latitude of 65° N, the increased insolation during the next year may be found at 63° N of the present time. Such curves were calculated by Milankovitch twice. The curves published in the work of Köppen-Wegener were based on Stockwell-Pilgrim data, and those published in 1930 and 1938 on Leverrier-Michkovitch computations of the secular variations of astronomical elements for more precise values of masses and motion of other planets. One such curve was recently recomputed by Brouwer and van Woerkom.[6] The general trend in all these curves is essentially the same.

(4) The radiation curves can be drawn for the radiation corresponding to any time interval, but is seems that the most important for paleoclimatology are the curves representing the semiannual values of radiation. The term semiannual requires special attention. The difference between the astronomical summer and winter half year is at the present time equal to 7 days 14 hours. It could reach ±31 days 20 hours because of variations of astronomical elements in the geological past and, therefore, more precision must be used in the definition of a half year. The important concept of a "caloric half year" introduced by Milankovitch refined the theory at this point. According to the definition, the tropical year is divided into two numerical equal parts, so that during every day of the "summer caloric half year" the radiation at a latitude is greater than during any day of the "winter caloric half year." The theory is very sensitive at this point, as is proved by the fact that Hargreaves, using annual radiations instead of semiannual, could obtain only small variations of temperature (see Simpson[7]). In this connection the definition of a "caloric equator" also should be mentioned. The caloric equator is determined by the condition $Q_s = Q_w$, that is, by the equality of the summer and winter semiannual radiation.

(5) The curves calculated first by Milankovitch in 1924 showed a very irregular character of fluctuations of solar radiation. They were constructed from a set of radiation values for 5000-year-long intervals and are similar for various latitudes. One of the most remarkable features was a certain parallelism between the summer radiation curves and the succession of glacial and interglacial phases in the Pleistocene. The irregular shape of radiation curves is

due to a very complicated interplay of variations of astronomical elements. In spite of this irregularity there are groups of lowest minima that are close in timing to glacial phases. This fact alone leads to the conclusion that the fluctuation in radiation due to superposition of periodic variations of three elements of the earth's orbit have caused the corresponding changes in glaciation.

However it must be pointed out that the investigations of Milankovitch had much more general purpose than to give a theory of ice ages.

After the problem of solar radiation (that is, of insolation) was solved by exact methods in the astronomical part of the theory, the next step was to show how the variations of radiation are transformed into temperature variations, namely to provide a link between the solar radiation and the thermal phenomena on the earth. This link must be based on several physical laws.

(6) In this second or physical part of the theory, a chain of problems is involved in which meteorological and geophysical factors must be introduced. If we consider, for example, the troposphere, we know that the thermal effects of solar radiation depend on those factors in an extremely complicated manner.

In order to obtain a relationship between the radiation and the temperature of the lowest air layer in mathematical form, many simplifications are necessary. The derivation of a formula that will cover all important geophysical factors is reserved for the future. However, Milankovitch succeeded in deriving a formula that accounts for the effect of several physical factors. If θ is the absolute temperature, W is the radiation at a given latitude, A is the reflection coefficient of a unit area of the earth's surface covered by the atmosphere, M is the mass of the air column, k is the absorption coefficient for long waves, and σ is the constant factor in Stefan's law, the important relationship given by Milankovitch is as follows:

$$\sigma[\theta(0)]^4 = \tfrac{1}{2}(1 - A)(1 - kM)\, W \tag{1}$$

where $\theta(0)$ is the temperature of the lowest layer of the atmosphere. At this stage of approximation a "solar temperature distribution" is given for a steady state of radiation (insolation). No currents in the atmosphere and oceans are considered, and a certain average land distribution is postulated.

Because of their importance for further progress of this theory, I shall mention here basic physical assumptions. The problems discussed in this connection were as follows: heat conduction in the crust, heat exchange between the earth's surface and atmosphere, and the part played by water vapor in the atmosphere in the process of radiation (solar and earth).

The atmosphere was taken to be an ideal gas. The average cloudiness and composition were assumed to be functions of the distance from the ground. For the attenuation of light in the atmosphere, the Bougeur-Lambert's law was postulated and, for the dark radiation of the atmosphere the Kirchhoff-Stefan's law.

If in Equation **1** the absolute temperature is replaced by centigrade temperature u ($\theta = 273° + u$) and in the series expansion with respect to the ratio $u/273$ the higher powers than the first are neglected, **1** takes the form

$$u - u_0 = 173.8\,(W - W_0) \tag{2}$$

where u_0 and W_o are the values at the initial time. A similar formula for the summer caloric half year is

$$\Delta u_s = (1/150)\, \Delta Q_s \tag{3}$$

with the numerical coefficients corresponding to expressions previously used. Q is expressed in canonical units.

Equation **3** represents the necessary link between the temperature variations and the summer semiannual radiation.* Obviously this relationship is also important in the theory of ice ages.

(7) Terrestrial factors taken into account in derivation of Equations **1** to **3** were mentioned in previous sections. The theory can be generalized either by introducing new assumptions about those factors or by considering the part played by other astronomical or geophysical factors.

One important generalization was made by Milankovitch himself. In the version that leads to Equations **1** to **3** the reflection coefficient of the earth's surface (A) was assumed to be constant. In 1933 W. Wundt pointed out that during the time of a growing icecap the reflection of solar radiation will increase. Under certain conditions there will be almost no summer in the polar regions. According to Wundt the reflection in the Northern Hemisphere could increase by about 3 per cent. This would result in an essentially lower average temperature. After J. Devaux published in 1933 the results of his observations concerning the reflection of radiation by snow in the Alps and other regions, Milankovitch introduced into the theory a variable reflection coefficient. He was then able to show that this factor makes the minima of radiation curves deeper. Thus an increasing reflection coefficient can intensify the process of growing of icecaps† and the lowering of the snow line. The problem of variations of the altitude of the snow line was discussed in detail by Milankovitch in his work in 1941, who showed that the lowering of the snow line in mountains by 1.094 m. corresponds to a decrease of summer radiation by 1 canonic unit. Thus the elevation of the snow line could decrease several times during the last 600,000 years by about 1000 m., increasing again during the interglacial phases.

The final results of Milankovitch's calculations are presented in FIGURE 1 and in TABLE 1.

(8) In conclusion I add only the following remarks. Several facts mentioned above can be incorporated in a theory of ice ages. The parallelism between the lowering of the snow line in the Alps, the equatorial Andes, and equatorial East Africa, and growing continental ice sheets was long since noted. The existence of groups of minima of summer radiation separated by time intervals with higher radiation appears to be similar to a succession of glacial and interglacial phases established by geological methods. This fact hardly can be called accidental (see Zeuner,[8] pp. 142 and 143). In the opinion of many investigators it implies a causal connection.

There can be no doubt that a lower summer radiation should cause the

* The parallelism between temperature variations and those of the summer semiannual radiation defined by Equation **3** is caused by the fact that the first approximation was used in deriving this equation. According to Equation **1** this link is more complicated.

† A growing icecap was the primary effect; its cooling action was a secondary effect.

growth of icecaps and the lowering of the snow line. The question is only to what extent. Moreover a varying reflection coefficient could cause delay of a temperature minimum for many years with respect to the time of the lowest radiation. A second factor could increase the delay to the order of thousands of years, namely, the melting of large ice masses. Other factors could also contribute to the generation of ice ages, such as the lowering of the sea level due to accumulation of ice and snow masses in some regions of the earth; re-

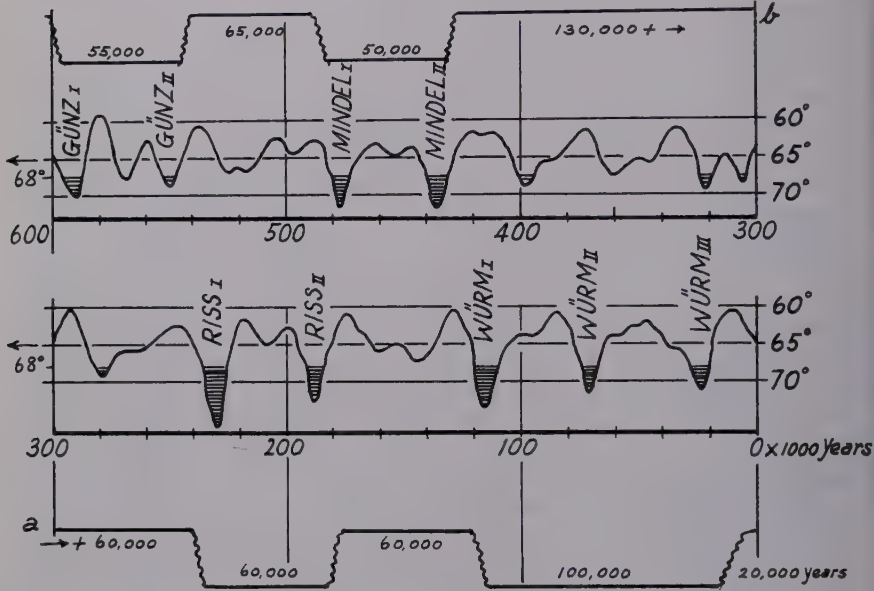


FIGURE 1. Secular variations of the summer radiation at 65° N represented by latitude oscillation (Milankovitch[5]). The line *a-b* is representing the rhythm of 4 ice ages (without differentiation).

TABLE 1

Southern Hemisphere				Northern Hemisphere			
Time in thousands of years	Snow line: lower by m.	Average temperature, summer half year decrease (°C.)	Annual temperature decrease by: (°C.)	Time in thousands of years	Snow line: lower by m.	Average temperature, summer half year decrease (°C.)	Annual temperature decrease by: (°C.)
600	910	6.1	1.2	590.3	950	6.3	2.2
560	1030	6.9	1.1	550	700	4.7	1.5
485	830	5.5	0.8	475.6	1070	7.1	2.5
444	840	5.6	0.9	435	1140	7.6	3.1
313.4	980	6.5	1.2	230	1540	10.3	4.0
198.5	930	6.2	0.9	187.5	1190	7.9	2.8
152.2	840	5.6	1.0	115	1320	8.8	3.3
105.1	1110	7.4	1.4	71.9	1100	7.3	2.8
30	830	5.5	1.3	25	1010	6.7	3.0

distribution of masses in general, especially masses in the hydrosphere, in oceanic currents, and elsewhere; and intensified anticyclonal conditions due again to the accumulation of ice just mentioned. A quantitative discussion of all these factors is still waiting its turn, thus leaving room to different theories of ice ages.

It is evident, however, from the review of the investigations of Milankovitch that the results of his theory form a solid base for the mathematical approach to paleoclimatic problems. In this theory all assumptions are connected with well-established physical laws. Like many other theories this one is certainly open to many generalizations.

A theory of ice ages that constitutes only one of the applications of the theory of thermal phenomena is not the point that can prove or disprove the validity of the investigations of Milankovitch, which will remain a very important contribution to the mathematical theory of the secular variations of climate independent of the phenomenon of ice ages.

As to the phenomenon of ice ages, a better understanding can be created, in my opinion, only by a unified theory in which it will be shown by mathematical methods how several factors now appearing to be of secondary importance contributed to the processes of glaciation. At the present time we have an exact theory of the variations of solar radiation caused by variations of astronomical elements of the earth's orbit. The physical part of this theory implies that these variations undoubtedly favored a sequence of glacial phases in the Pleistocene. However, it seems to be necessary not to make just a single factor responsible for ice ages in all geologic eras as has been done in some theories of this complicated phenomenon.

An important place in a unified theory is reserved for the polar wandering and, probably, for the displacement of continents.

References

1. MEECH, L. W. 1857. On the relative intensity of the heat and light of the sun upon different latitudes of the earth. Smithsonian Contrib. to Knowledge. **IX**. Washington, D.C.
2. MILANKOVITCH, M. 1920. Théorie mathématique des phénomènes thermiques produits par la radiation solaire. Gauthier-Villars. Paris, France.
3. MILANKOVITCH, M. 1930. Mathematische Klimalehre und Astronomische Theorie der Klimaschwankungen. I. Part. A. *In* Köppen-Geiger's Handbuch der Klimatologie. Berlin, Germany.
4. MILANKOVITCH, M. 1933. Astronomische Mittel zur Erforschung der erdgeschichtlichen Klimate. IX. Part VII. *In* B. Gutenberg's Handbuch der Geophysik. Berlin, Germany.
5. MILANKOVITCH, M. 1941. Kanon der Erdbestrahlung und seine Anwendung auf das Eiszeitenproblem. Ed. spec. Acad. Royale Serbe. CXXXII. Belgrade.
6. VAN WOERKOM, A. J. J. 1953. The astronomical theory of climate changes. *In* H. Shapley, Ed. Climatic Change Evidence, Causes and Effects. Harvard Univ. Press. Cambridge, Mass.
7. SIMPSON, G. C. 1940. Possible causes of change of climate and their limitations. Proc. Linnean Soc. **152:** 190–219.
8. ZEUNER, F. E. 1952. Dating the Past. Methuen. London, England.

SOLAR-TERRESTRIAL CLIMATIC PATTERNS IN VARVED SEDIMENTS*

Roger Y. Anderson
University of New Mexico, Albuquerque, N. Mex.

Introduction

Joseph Barrell (1917) in speaking of a thinly laminated anhydrite said: "Such a series of beds is worthy of very careful analytical study as a detailed meteorological record of the past." Barrell was referring to the principle that the thickness of an annual sedimentary layer is a function of climate. It is theoretically possible to translate climatic changes into sediment layers of varying thickness and, if there is a solar effect on climate, it should be recorded in varves. This influence might be demonstrated in several ways, and this paper presents the evidence compiled to date.

Seasonal Interpretation

It is fair to state at the outset that there has not always been agreement that thinly laminated sedimentary deposits represent seasonal deposition. The most impressive evidence has come from studies of recent sediments where varves are in the process of formation. At least 35 examples have been reported from Europe and the Near East where most of the work has been done: 5 from North America and 3 from the Far East. The generalizations derived from the study of these deposits show that there is one outstanding feature common to nearly all of them. This is the deposition of some sort of an organic layer each year (Bradley, 1937, p. 32). This may be a well-defined layer of aquatic algal ooze or may be represented by only a gradual transition from a light to dark color. Inorganic glacial varves are an exception, but virtually all of the recent clastic varves and all but one (?) of the recent evaporite sequences including halite have this feature. The organic layer represents a yearly interval of high growth and mortality in the aquatic population; once the bottom water of a lake, lagoon, fjord, or bay becomes stagnant and unventilated, it takes only the addition of a slight amount of clastic material or chemical precipitate to form a recognizable varve. The most important complication of the previous generalization is caused by changes in the growth and mortality of aquatic organisms. There may be several such intervals in one year, although this factor does not seem too important, and the supernumerary layer can often be recognized as such. More confusing are diatom-organic laminae where diatom flowering and deposition are coincident with organic deposition. In this case, several periods of flowering may result in a multiplicity of layers each year (Deevey, 1953, p. 291).

If the organic-layer criterion is applied to ancient varves it is found that most of the reported examples do represent seasonal deposition. The greatest exceptions are iron-stain laminations, where the Liesegang mechanism (Stetson, 1933; Sugawara, 1934) is capable of producing varvelike layers and alternating

* This work was supported in part by Research Grant NSF-G17144 from the Earth Sciences Program of the National Science Foundation, Washington, D.C.

sand, silt, and clay laminae without an organic fraction. Also the strong seasonal control on sedimentation exerted by the freeze and thaw of ice can form glacial varves without a well-defined organic layer. A list of varved deposits would be too long to be included here, but it is safe to say that every geologic period from each continent has many examples. These would include varves whose second component consisted of sand, silt, clay, calcite, dolomite, anhydrite, halite, diatoms, and calcitic shell material. In addition some iron, potash salt, bittern, and silica laminae might also be included.

Climatic Interpretation

Varve thickness can be an approximate index of precipitation and temperature. Other factors such as evaporation, storminess, and insolation also enter in but probably cannot be treated separately in ancient varves. Unfortunately, even the effects of temperature and precipitation are difficult to separate, but estimates can be made after a study of the varve type and the environmental setting of the basin. For example, thickness changes in the clastic layer of organic-silt laminae would most likely represent changes in runoff and, indirectly, precipitation. In special cases pollen analysis of microtome sections of individual laminae can reveal the season of greatest runoff and, perhaps, the type of precipitation (Welton, 1944; Anderson, 1961). The thickness of a glacial varve is thought to be an index of glacial melting and, indirectly, of temperature. The interpretation of calcite varves is difficult because thickness changes may be the result of the dilution of basin water by precipitation, or they may be caused by changes in CO_2 content arising from variations in water temperature, evaporation, or organic activity. In the case of evaporitic anhydrite laminae forming in a lagoonal basin with little interior drainage, the effects of temperature and evaporation would overshadow those of precipitation. Obviously the problems of interpretation are complex, and each basin must be considered separately in order to determine to what extent temperature, precipitation, or both are being measured by varve-thickness changes.

Comparison of recent varves with meteorologic observations will be necessary before the influence of climatic factors on varve deposition can be fully known. Little work has been done. Shostakovich (1931, 1936) presented data relating varve thickness in Pert Lake, Finland, and Saki Lake, Crimea, to local precipitation. He found that a twofold increase in seasonal precipitation brought about nearly the same increase in varve thickness. He also found a close relationship between the thickest varves and very wet years and between the thinnest varves and very dry years. In a different type of study, Granar (1956) related varve thickness in Nyland Fjord, Sweden, to discharge records for associated rivers. For some intervals the correlation he obtained was weak, but for others the correspondence was outstanding. Late-glacial and postglacial varve correlations by DeGeer and Antevs in Europe and North America are another point of evidence. Their success is indirect testimony that the climatic changes recorded in varves are sufficiently strong and consistent to be traceable over a wide area.

One can state that there is a relationship between seasonal weather and varve thickness, and preliminary studies indicate that it is an important one.

Evidence of a Solar Influence on Terrestrial Climate

Recognition of Solar Periods

One way of demonstrating a solar-climatic influence is to recognize the 11- or 22-year solar cycle* in varve data. Meteorologists seem divided on the question of a solar period in the data from the earth's lower atmosphere. This in itself suggests that if there is an effect it is slight, and might appear in the varve record only under favorable circumstances. So many investigators have found evidence for an 11-year period in meteorologic data that a solar influence would seem a certainty (Nupen and Kageorge, 1958, p. 5). Other meteorologists argue that the statistical techniques used in these studies were of doubtful validity and that no good pertinent evidence has ever been established (F. W. Ward, Jr., personal communication, 1959).

The number of reported occurrences of the 11- and 22-year period in varved sediments is also impressive (FIGURE 1). The same argument of inadequate statistical technique could also be used against the evidence. In general, analysis has been either graphic-visual, or has employed some form of harmonic analysis. Graphic plots of published varve series leads one to suspect that many of the reported occurrences of the 11-year solar period are probably erroneous. Some are merely averages between prominent maxima that are a great deal more variable than the 11-year sunspot maxima intervals (Bradley, 1929, FIGURE 15). In most cases this variable interval between prominent maxima seems to average about 5 years. However, some of the curves published by Korn (1938, FIGURES 14 and 15) show a remarkably uniform interval at about 11 years, and parts of other series show isolated groups of maxima about 11 years apart that could represent the solar period (Anderson and Kirkland, 1960, FIGURE 9, units near 500).

Shostakovich (1944) analyzed cores from 19 recent lakes in Europe and Siberia (FIGURE 1). He found remarkably close agreement between 3 periods that average 2.7, 5.6, and 10.8 years. Brooks (1928) observed 2.15- and 5.1-year periods in glacial varves. He also recognized 2+- and 5+-year periods in modern meteorologic data. The 11-year period has generally been the most prominent and persistent one when several periods have been found in a varve series, except in some glacial varves where the weak "dearth cycle" of about 10.4 years has been reported (Brooks, 1928). In spite of these few consistencies for periods of 2+ and 5+ years, the most outstanding feature of the summary of reported periodicities (FIGURE 1) is the lack of general agreement for periods other than 11 years. This probably reflects the diversity of techniques employed in the study of cycles, and the fact that periodic tendencies, when present, are weak, the result of chance, or occur only under special local conditions.

Recent refinements in statistical theory and the development of electronic

* In this paper a cycle is considered any kind of phenomenon that tends to return to an original position, status, or value. Cycles may recur in a predictable sequence but need not appear at a regular time. A period or rhythm is considered a cycle that recurs at a predictable time. For example the waxing and waning of sunspot frequency is a cycle. The alternation of large and small maxima is another cycle, but the recurrence of maxima at approximate 11- and 22-year intervals is a period or rhythm.

computors have made possible a nearly objective technique for the analysis of time series for periodicity. This "power spectrum" method involves Fourier analysis and autocorrelation, and has become a widely accepted technique for separating "signals" from "noise" in time series of rapidly fluctuating elements

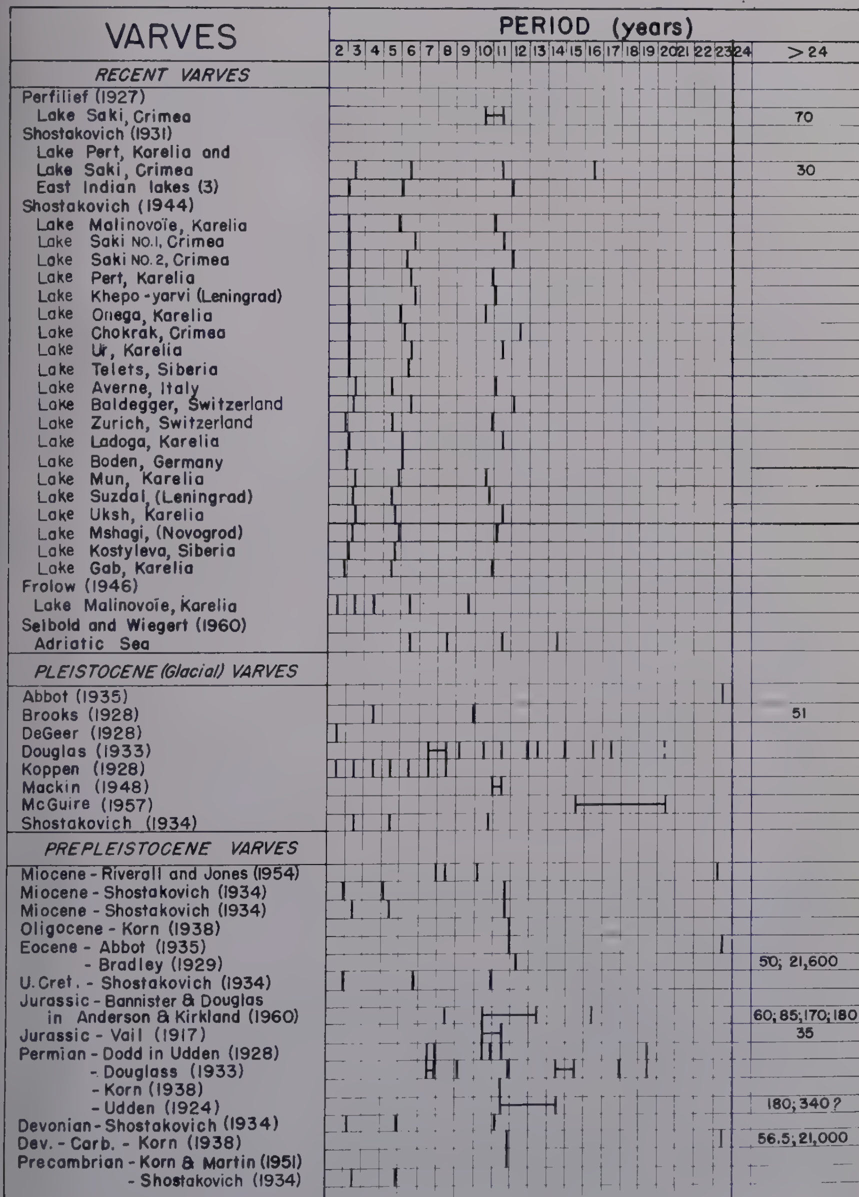


FIGURE 1. Summary of reported periodicities in varve series.

(Landsberg *et al.*, 1959). The method has been adapted for use on meteorologic data by Ward (1957) and Landsberg *et al.* (1959), and is described fully in their articles, and by Blackman and Tukey (1958). Landsberg and his co-workers analyzed a homogeneous series of temperature data from a recording station in Maryland and found an 11-year period corresponding to the sunspot period. The effect was slight, but suggests that changes accompanying the 11-year solar period are capable of being impressed on the lower atmosphere and might appear in varves under certain conditions. Landsberg and his colleagues also found weak periods near 2 and 5.6 years which is in general agreement with the results of Brooks (1928) and Shostakovich (1931, 1934, 1944) (FIGURE 1).

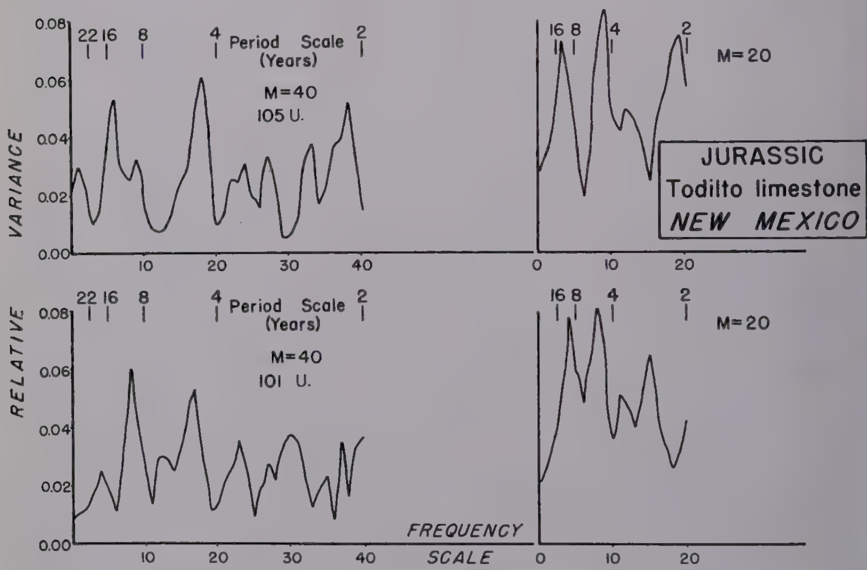


FIGURE 2. Power spectra of Upper Jurassic Todilto limestone varves, New Mexico.

Power spectra of several varve series are available. Two short series of 105 and 101 U. from the Jurassic Todilto limestone of New Mexico were analyzed (FIGURE 2). The Todilto limestone varve-thickness changes are thought to reflect both temperature and precipitation, although the effect of temperature is probably greater (Anderson and Kirkland, 1960). According to Ward (personal communication, 1959), ". . . only one point on both graphs of the varve spectra is very much above average, at a frequency of 17. This corresponds to a period of 4½ to 5 years. It would seem that by chance there would be one like this. There is little evidence of a sunspot cycle period." Unfortunately, interpretation of the spectra is somewhat subjective where the effect is slight, although confidence estimates can be established. The spectra agree in a general way with the findings of Landsberg *et al.* (1959). Both the Todilto and the Maryland temperature record have peaks near 2 and 5 years, and the 10- and 13-year Todilto peaks occur near the sunspot period. The

analysis of many series would be needed to demonstrate a conclusive relationship with periodic tendencies of this magnitude.

The most interesting series analyzed by Ward is a 470-U. sequence from a core of the Upper Devonian Ireton shale member of the Woodbend formation, Alberta, Canada. The laminations are made up of couplets of green-gray calcareous shale and brown organic-rich calcareous shale that average slightly under 2 mm. in thickness. The varves are marine, and were formed as a calcareous ooze with both a terrigenous and chemically precipitated component. Temperature exerted some control over the formation of the chemical precipitate. The controls for the clastic component might be oceanic circula-

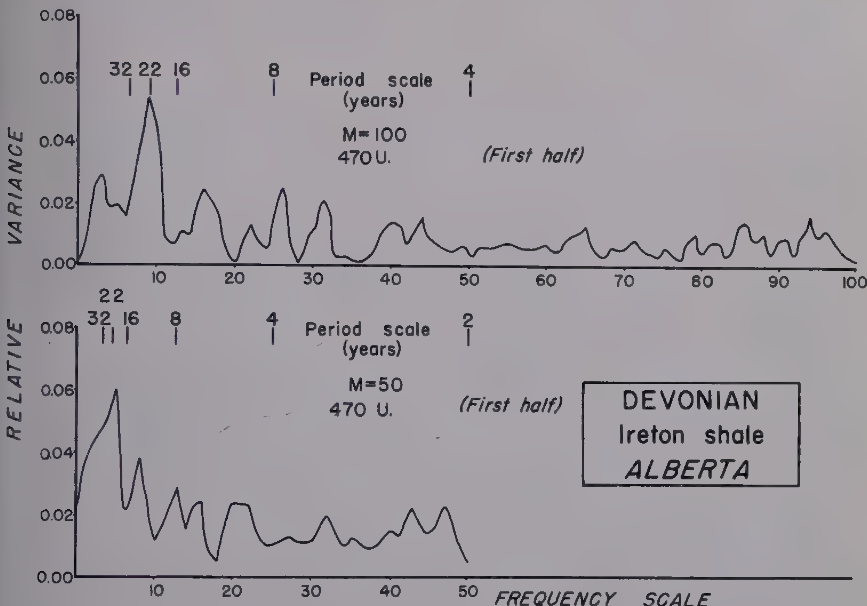


FIGURE 3. Power spectra of the Upper Devonian Ireton shale varves, marine Woodbend formation, Alberta, Canada (*first half*).

tion or storminess and precipitation. The power spectra from the Ireton shale show a prominent peak at the frequency 9 (lag = 100) corresponding to a period of 22 years (FIGURES 3 and 4). Ward (personal communication, 1960) believes that the peak is significant, but he would reserve judgment until a longer series can be analyzed from either end of the measured interval, or until an independent series is available. The series, however, is a long one in so far as meteorologic records are concerned. Spectra computed for each half show a peak at the same point, indicating that the period was present throughout the data and increasing the probability that the period has physical reality. The 22-year period is not obvious in a simple graphic plot of the varve-thickness changes but apparently is responsible for the group of prominent maxima in the smoothed curve of FIGURE 7. These maxima show a remarkably uniform recurrence at little more than 21 U. apart that closely resembles the

spacing of the 22-year cycles in the modern sunspot record (compare with FIGURE 6).

Tentative results from harmonic analyses of several other series are available. The analyses were done by L. H. Koopmans of the Sandia Corporation, Albuquerque, N. M., using an IBM-704 computor and programs modified from those originally developed by Tukey. This brings the total number of spectra available to 12 as follows.

Analyzed by Ward (1957): Ireton shale, Upper Devonian, 470 U.; Todilto limestone, Upper Jurassic, 105 and 101 U.; and Puente formation, Miocene, 480 U.

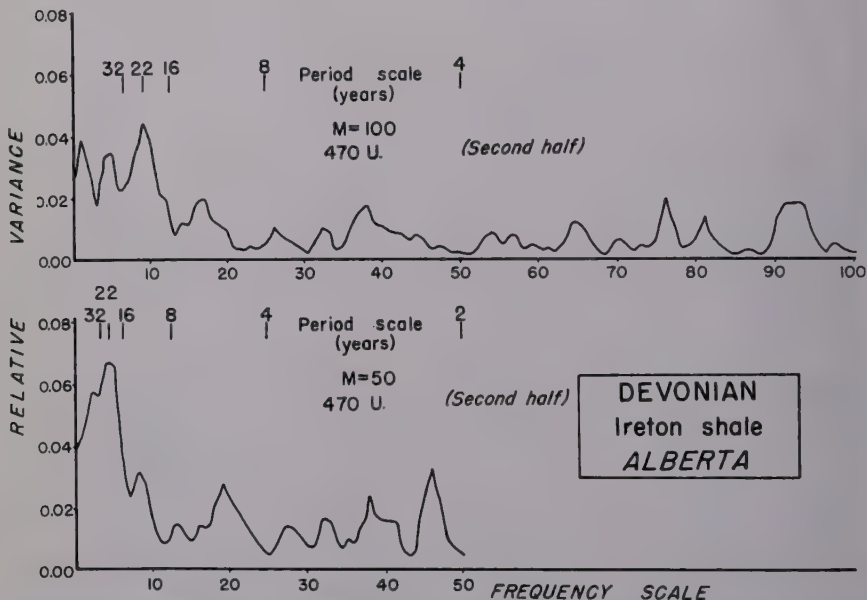


FIGURE 4. Power spectra of the Upper Devonian Ireton shale varves, marine Woodbend formation, Alberta, Canada (*second half*).

Analyzed by Koopmans (tentative): Todilto limestone, Upper Jurassic, 943 U.; Nyland Fjord, recent, 100 U.; Averne, recent, 157 U.; Chokrak, recent, 76 U.; Lake Onega, recent, 81 U.; Lake Saki, recent, 145 U.; Lake Teletz, recent, 50 U.; and Lake Ur, recent, 97 U.

Published sources referring to the above series are as follows: Todilto limestone, Anderson and Kirkland (1960); Puente formation, Riveroll and Jones (1954); Nyland Fjord, Granar (1956); Lakes Averne, Chokrak, Onega, Saki, Teletz, and Ur, Shostakovich (1934).

As might be expected, many of the peaks in the spectra of the above series are weak, and many are probably chance variations that do not reflect consistencies in the data. In order to facilitate interpretation, all the prominent peaks in the above spectra (except the short Todilto series) were plotted and the results summarized (FIGURE 5). The strongest peak in each series appears

as a longer bar on the graph; peaks whose periods were less than about one fourth the length of the series were not plotted. Little is known of the technique for gathering some of the data. No attempt will be made here to evaluate climatic significance or data reliability and the 10 varve series must be assumed to be representative. If the peaks are randomly distributed there should be a gradual increase in the number of peaks toward the right of the chart.

Interpretation must be subjective, and some of the series are so short that the significance of the position of some of the peaks at the longer periods is low. The 11-year period is not present. Instead the frequencies near 12 to 14 years and 8 to 9 years seem better developed. This distribution could be due to chance, but if it persists as more series are analyzed one might expect a causal relationship. A Fourier analysis of recent varves from the Adriatic

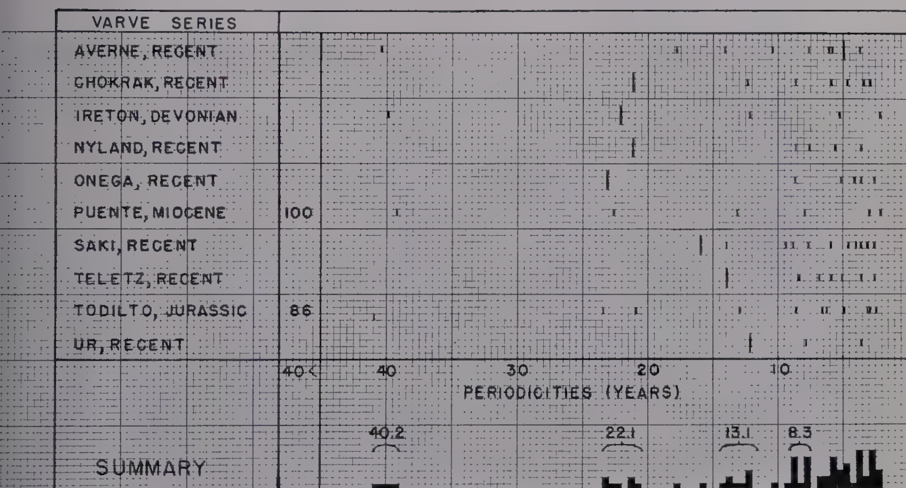


FIGURE 5. Summary of prominent peaks in spectra from 10 varve series.

area made by Seibold and Wiegert (1960) found tendencies near 6, 8, and 14 years throughout an 820-U. record that agree quite closely with the summary in FIGURE 5. They also found a weak 11-year period in their data. The complex of peaks between 2 and 6 years either represents "noise" or some factor of local climatic persistence that might account for the interpretation of periods between 2 and 6 years by Brooks (1928), Shostakovich (1931, 1934, 1944) (FIGURE 1), and Landsberg *et al.* (1959).

The appearance of several peaks near 22 years, the prominence of some of them, and the slight separation from adjacent peaks support the earlier supposition that the 22-year solar cycle has a slight influence on climate. One might speculate as to why a 22-year period and not the 11-year period should manifest itself in the varves. Since about 1850 the Zurich mean annual relative sunspot numbers show the 22-year solar cycle as higher peaks on alternate 11-year maxima. One estimate fixes the difference at about 22 per cent (Nicholson, 1929, p. 80). There is also a difference in the spacing of maxima

in alternate cycles. The 11-year influence, modified by terrestrial factors, might be so weakened or modified that the 22-year period, with a longer interval of climatic adjustment, could appear as the stronger of the 2 solar cycles. Alternatively the 22-year magnetic cycle of sunspots may exert a disproportionate control on weather and climate.

Meteorological series are too short for the study of the 22-year period over long intervals, and further work with varves and tree rings will probably yield the most conclusive results. In this respect it may be important that the examination of many long tree-ring series by Schulman (1945, 1956) led him to believe that the cycle near 22+ years was the strongest recurring phenomena in the general tree-ring index.

Comparison of Long-Term Trends

Trends between 50 and 500 years. With some evidence for a physical relationship between solar and climatic changes at the 22-year level it is desirable to compare trends that are somewhat longer. The 211-year record of sunspot data is the most reliable measurement available for comparison, and other solar phenomena can be related to it. In order to obtain a graphic idea of the observed changes greater than 30 years in length, the Zurich mean annual relative sunspot numbers were averaged over an arbitrary interval of 31 years and plotted (FIGURE 6). The curve is shorter than 211 years because 15 U. at each end were lost by the smoothing operation. The most obvious character of the curve is the significant change over a long period of time. The intervals between the 2 maxima are 68 years and more than 100 years. The interval between minima is about 86 years. The average number for the decade ca. 1775 is about 64, dropping rapidly to only 22 in the decade ca. 1810. A lengthening of the interval between 11-year maxima accompanied this strong long-term minimum. A less intense maximum and minimum occurred near 1850 and 1900 respectively, and the present maximum has already exceeded the one near 1775. Assuming that other solar phenomena are related to changes in sunspot activity, it is apparent that large oscillations on the order of 70 to more than 100 years are taking place within the observable record. The differences between long-term maxima and minima should be sufficient to be reflected in meteorologic observations and also in varved sequences if there is any effect at all on the 22-year level.

When the smoothed sunspot curve is compared to a similarly constructed curve from the Upper Devonian Ireton shale the result is impressive (FIGURE 7). The Ireton trends are of nearly the same duration as those of the observed sunspots. The 22-year period, already indicated in the power spectrum, is most obvious between units 200 and 340 and was strong enough to appear after the 31-year averaging. Furthermore the association of the prominent 22-year maxima in the Ireton sequence with the gradual buildup of an unusually strong maxima near unit 330 suggests that solar activity played some direct role in intensifying that trend. In order to obtain a graphic estimate of the amount of oscillation that might be due to random-thickness changes, thickness measurements were selected randomly from a pool of values and plotted, using the same smoothing formula. If nature had selected varve

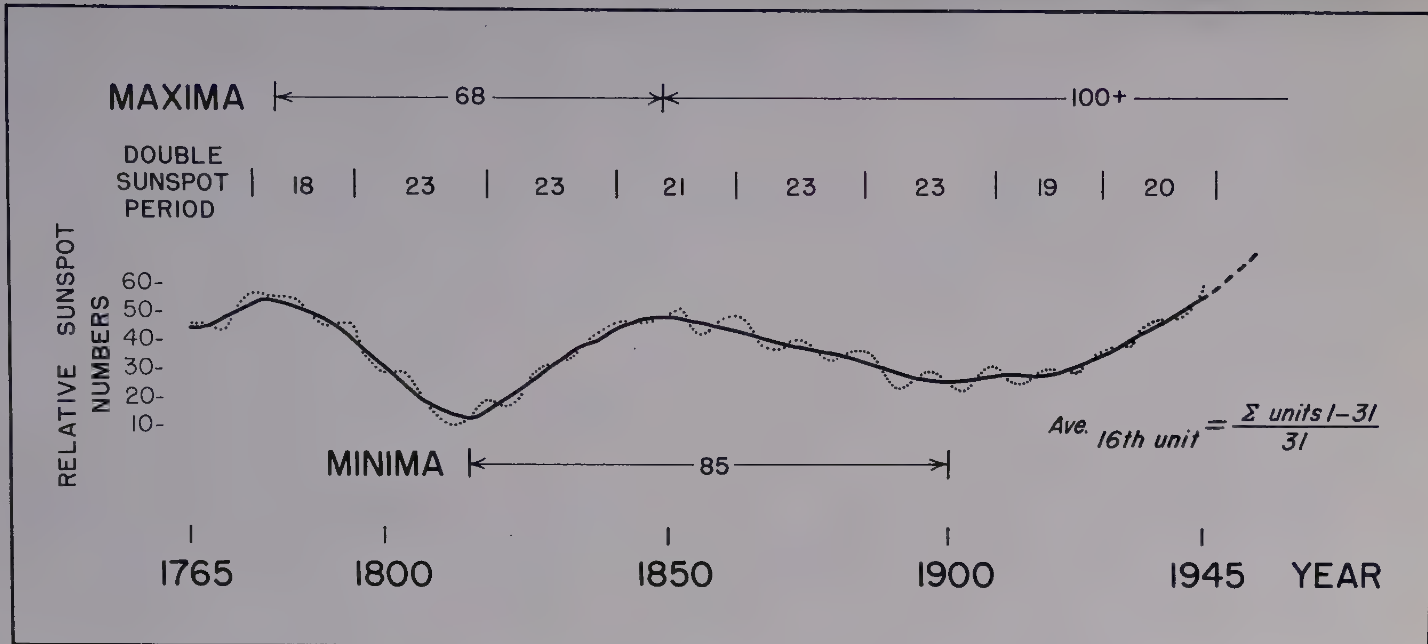


FIGURE 6. Smoothed curve of mean annual relative sunspot numbers (Zurich, Switzerland).

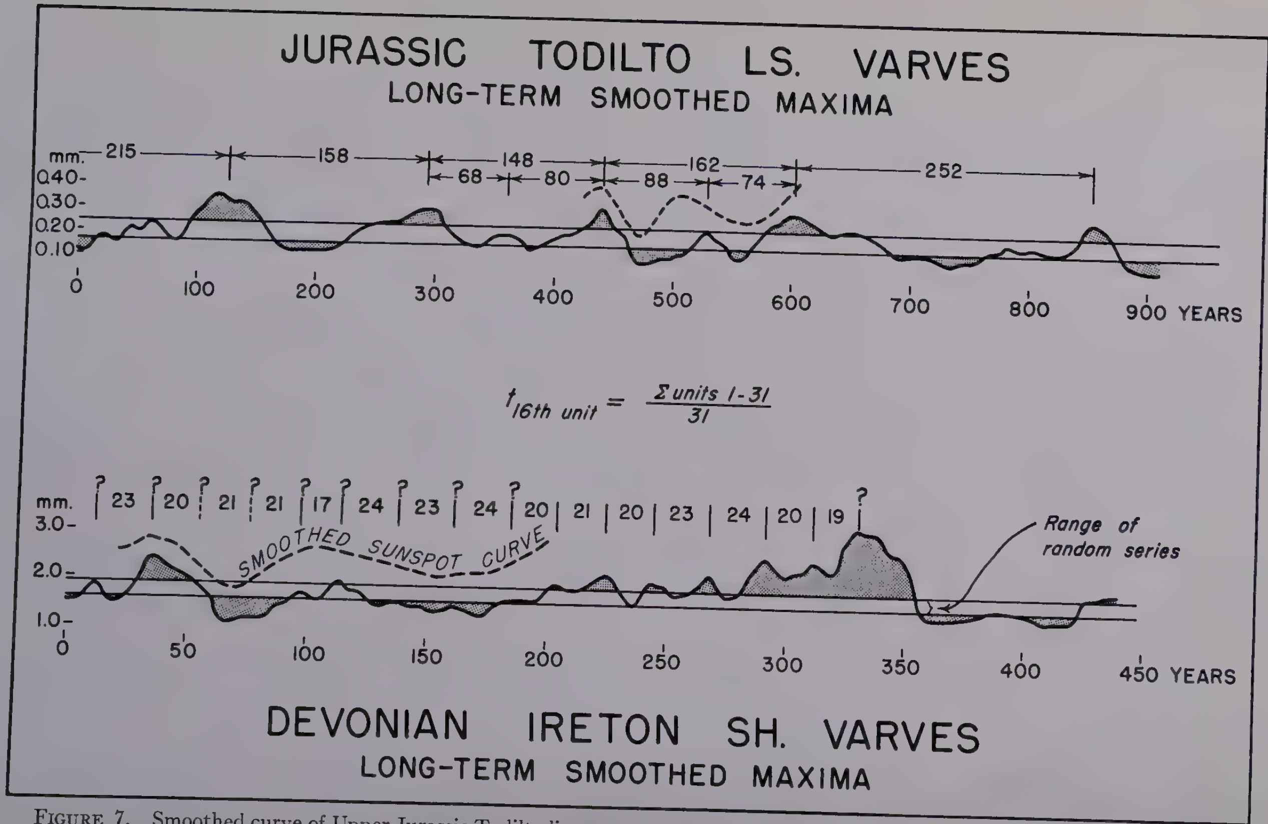


FIGURE 7. Smoothed curve of Upper Jurassic Todilto limestone varves, New Mexico, and the Upper Devonian Ireton shale varves, Alberta, Canada.

thickness individually, in a completely random manner, the natural and artificially selected curves should be similar. The random curve, however, is monotonous and shows none of the extreme variations or long-term trends found in the natural sequence. The extreme range of the randomly selected values is compared to the natural sequence in FIGURE 7.

A similar plot was made for 943 varves from the Jurassic Todilto limestone (FIGURE 7). The intervals between maxima that extend above the range of random values are between 148 and 252 years, and average 187 years. Smaller maxima appear between the larger ones and display a pattern of 68 to 88 years between maxima. These shorter trends resemble the smoothed sunspot curve even more closely than those in the Ireton sequence, and may be responsible for the strong 86-year peak in the harmonic analysis of the Todilto limestone by Koopmans (FIGURE 5). These quasi-periodic trends probably represent long-term persistence in temperature, precipitation and, perhaps, oceanic circulation, and their similarity to observed sunspot trends suggests a relationship.

Trends greater than 500 years. The 1592-year Todilto varve record (Anderson and Kirkland, 1960, figure 9) shows cycles on the order of 180 years through only about two thirds of the sequence. Both ends of the series have closer and weaker maxima. These differences may be the result of changes in the pattern on the order of 1000 years although a much longer sequence would be needed to see if the 86- or 180-year quasi-period would recur.

The physical effects of climatic trends on the order of 2000 to 5000 years are largely unknown. Late-glacial and postglacial pollen and varve studies have shown that important changes are taking place (Willett, 1953, p. 55), but a solar control could only be inferred. The late-glacial and postglacial varve record is not complete or homogeneous enough for detailed analysis, although cores from recent lakes may eventually accomplish this. The ultimate cause of many kinds of stratification is still a geologic mystery. Climatic changes on the order of 5000 years are a reasonable interpretation (Barrell, 1917), but a solar control is not demonstrable.

The climatic change accompanying the 21,000-year equinox cycle is apparently capable of causing differences in lithology. Good evidence is found in the Eocene Green River beds of Wyoming where the varves permit the calibration of deposition rates for numerous alternations of oil shale, and marlstone (Bradley, 1929, p. 105). The average time involved in the formation of an oil shale-marlstone couplet was 21,630 years. A similar conclusion had been reached by Gilbert (1895) for 18-inch limestone-shale alternations in Cretaceous sediments. One interesting observation is that the changes in lithology occur suddenly although the trends that caused them were probably gradual. Many gross changes in lithology are probably the result of climatic trends of several thousand years duration, but this would be difficult to demonstrate in the absence of a varve calibration.

Comparison with Known Solar Changes

The similarity between the Ireton (Devonian)-Todilto (Jurassic) cycles and the observed sunspot changes suggests solar control, but the sunspot record

is so short that it will be difficult to demonstrate. Correlation of varve changes in recent lakes with the sunspot record offers some hope of relating the two. Shostakovich (1931, 1944) compared the thickness of recent varves in Pert Lake, Karelia, Saki Lake, Crimea, and Malinovoïe Lake, Karelia, with 11-year sunspot maxima since 1738, and found essential agreement with an average difference of only 0.5 years between sunspot maxima and mud-thickness maxima. The power spectrum for Saki Lake (FIGURE 5) would not seem to support this conclusion however. Shostakovich did not consider long-term changes but if his smoothed curve from Pert Lake is compared to the graph of mean annual relative sunspot numbers there is fair agreement between the maximum and minimum near 1775 and 1810 (FIGURE 8). The Pert Lake curve should reverse its trend after 1850 but continues to a maximum thickness. Shostakovich made no adjustment for lack of compaction in the upper part of the sequence but believed it could be a factor. This lack of compaction might account for part of the discrepancy. The Pert Lake curve has a strong minimum near 1700. The sunspot record is poor before 1749, but historical records show a great dearth in sunspots from 1645 to 1715 (Stetson, 1947, p. 166) that could correspond to the Pert Lake minimum near 1700. This means that 2 maxima and 1 minimum are in agreement with the possibility that the minimum near 1900 was obscured by lack of compaction of the upper layers of sediment.

No detailed attempt will be made to discuss the mechanisms connecting increased varve thickness with increased sunspot activity. Changes in ultraviolet and corpuscular radiation are known to accompany changes in sunspottedness. These changes are presumably capable of altering temperature and circulation patterns (precipitation) and, indirectly, varve thickness. The Pert Lake varves should be an index of precipitation, and thickness maxima appear to accompany long-term sunspot maxima and increased corpuscular radiation from the sun according to the reasoning of Willett (1953, p. 66). A similar relationship seems to hold for the short sequence from Nyland Fjord, Sweden, where the trend agrees with the minimum near 1900, and for the short series from Lake Malinovoïe, Karelia, which has a varve trend that agrees with the sunspot dearth near 1700 (FIGURE 8).

The smoothed curve from Saki Lake, Crimea, published by Shostakovich (1936) is quite different from that of Pert Lake. The maxima and minima have an almost opposite relationship in the last 400 years and there are more trend reversals. When compared to the graph of sunspot numbers there are about the same number of maxima, and the minima follow about 10 to 15 years after sunspot maxima.

The good fit of these curves could be estimated only after a careful study of all the factors of local and regional circulation. There is, however, some similarity between the number and spacing of trend reversals in the varve and sunspot records. The record of solar observations is so short that many recent varved deposits would have to be investigated to demonstrate a relationship conclusively. These early investigations do suggest, however, that more work along these lines would be justified, particularly in basins where temperature and not precipitation is the main controlling factor.

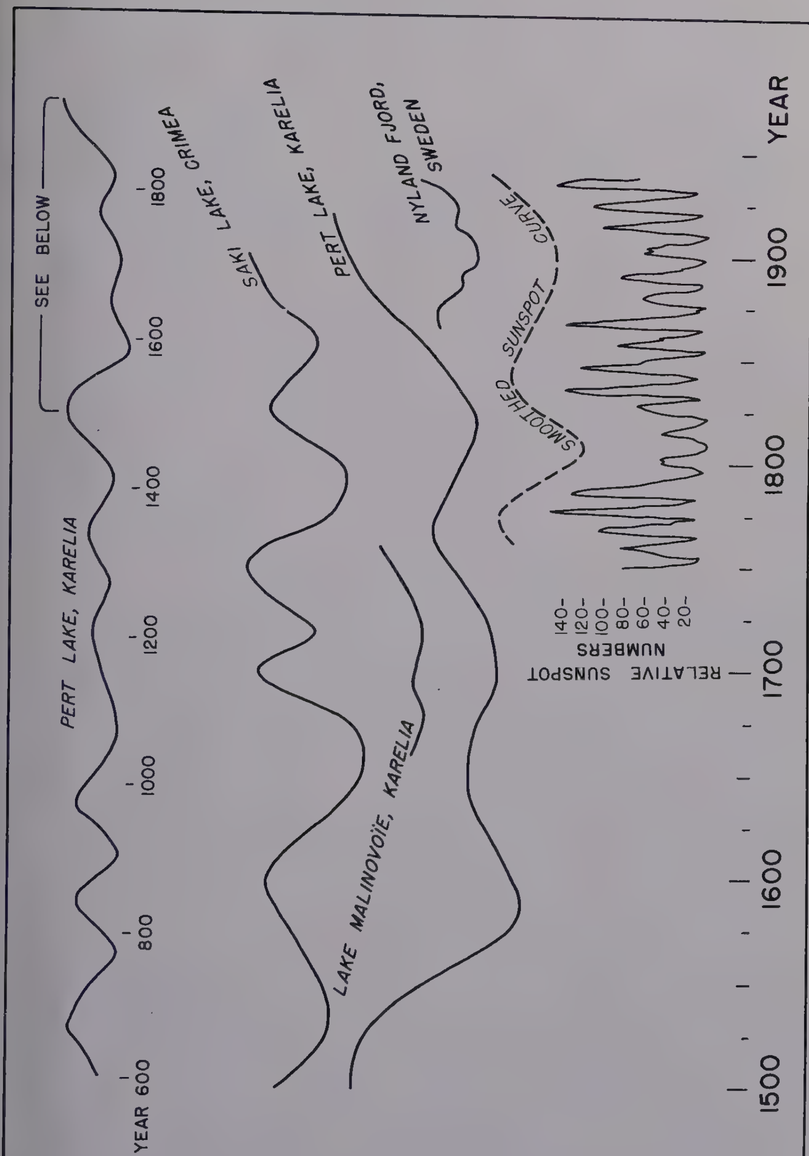


FIGURE 8. Comparison of long-term sunspot trends and varve thickness trends in recent lakes.

Conclusions

The 22-year solar period seems to be recorded in some varves, but the 11-year solar period may be modified by shorter-lived terrestrial factors so that it is unrecognizable or weak periodic tendencies of 12 to 14 and 7 to 9 years result. A complex of weak periodicities between 2 and 6 years may represent noise or local terrestrial climatic persistence. Longer varve trends with about 70 to 100 years between maxima may be related to similar long-term solar trends, and the varve record has revealed other quasi-periodic cycles of about double that length and of slightly greater magnitude. These observations comform to the generalization that both solar-climatic adjustment and total climatic effect increase with the length of oscillation.

Acknowledgments

I am grateful to F. W. Ward, Jr., of the Air Force Research Center, Cambridge, Mass., for analyses of the varve series shown in FIGURES 2, 3, and 4; to L. H. Koopmans of Sandia Corporation, Albuquerque, N. Mex., for the analyses represented in FIGURE 5; and to R. G. McCrossan of Imperial Oil Limited, Calgary, Alta., Canada, for the Ireton shale varve series.

References

Abbot, C. G. 1935. Solar radiation and weather studies. Smithsonian Misc. Coll. **94**(10): 89.

Anderson, R. Y. 1961. Evidence of seasonal lamination. Bull. Geol. Soc. Am. Abstr. In press.

Anderson, R. Y. & D. W. Kirkland. 1960. Origin, varves, and cycles of Jurassic Todilto formation, New Mexico. Bull. Am. Assoc. Petrol. Geol. **44**(1): 37–52.

Barrell, J. 1917. Rhythms and the measurements of geologic time. Bull. Geol. Soc. Am. **28:** 745–904.

Blackman, R. B. & J. W. Tukey. 1958. The measurement of power spectra from the point of view of communications engineering. Bell System Tech. J. **37:** 185–282, 485–569.

Bradley, W. H. 1929. The varves and climate of the Green River epoch. U. S. Geol. Survey Prof. Paper. **158:** 87–110.

Bradley, W. H. 1937. Non-glacial varves with selected bibiography. Nat. Research Council Ann. Report. App. A. Rept. Committee on Geologic Time. : 32–42.

Brooks, C. E. P. 1928. The problem of the varves. Quart. J. Roy. Meteorol. Soc. **54:** 64–70.

De Geer, G. 1928. Geochronology. Antiquity. **2:** 308.

Deevey, E. S. 1953. Paleolimnology. *In* Climatic Change, Harlow Shapely, Ed. : 273–318. Harvard Univ. Press. Cambridge, Mass.

Douglass, A. E. 1933. Tree growth and climatic cycles. Sci. Monthly. **37:** 481–495.

Frolow, V. 1946. Analyse de la série des varves du lac Malinovoïé. Acad. sci. Paris, compt. rend. **222:** 669–672.

Gilbert, G. K. 1895. Sedimentary measurement of Cretaceous time. J. Geol. **3:** 121–127.

Granar, L. 1956. Dating of recent fluvial sediments from the estuary of the Ångerman River (the period 1850–1950 A.D.). Geol. Fören. Stockholm, Förhandl. **78:** 654–658.

Köppen, V. 1928. Mehrjährige Temperatur schwankungen vor 8 bis 18 Jahrtausenden. Meteorol. Z. **45:** 263–265.

Korn, H. 1938. Schichtung und absolute Zeit. Neues Jahrb. Mineral. Geol. Paläont., Stuttgart. **74A:** 50–186.

Korn, H. & H. Martin. 1951. Cyclic sedimentation in varved sediments of the Nama system in South-West Africa. Trans. Geol. Soc. South Africa. **54:** 65–67.

Landsberg, H. E., J. M. Mitchell, Jr. & H. I. Crutcher. 1959. Power spectrum analysis of climatological data for Woodstock College, Maryland. Monthly Weather Rev. **87:** 283–298.

Mackin, J. H. 1948. Possible sun-spot cycle in pre-Wisconsin varves in the Puget area, Washington [abs.]. Bull. Geol. Soc. Am. **59:** 1376.

McGuire, R. H., Jr. 1957. A study of some Lake Missoula varves [Mont.]. Compass. **34:** 197–204.

NICHOLSON, S. B. 1929. Remarks on the sunspot cycle *in* Second Conference on Cycles. Carnegie Institute, Washington, D. C. : 79–80.

NUPEN, W. & M. KAGEORGE. 1958. Bibliography on solar-weather relationships. Amer. Met. Soc. Meteorological Abstracts and Bibliography, Malcolm Rigby, Editor. Washington, D. C. : 248.

PERFILIEF, B. W. 1927. Ten years of Soviet Science. Moscow. : 402, 403 (not seen, from Bradley, 1937 summary).

RIVEROLL, D. D. & B. C. JONES. 1954. Varves and foraminifera of a portion of the upper Puente formation (Upper Miocene), Puente, California. J. Paleontol. **28:** 121–131.

SHOSTAKOVICH, V. B. 1931. Die Bedeutung der Untersuchung der Bodenablagerungen der Seen für einige Fragen der Geophysik. Int. Ver. theor. angew. Limnol. Verh. Stuttgart. **5:** 307–317.

SHOSTAKOVICH, V. B. 1934. Mud deposits in lakes and the periodic fluctuation phenomena in nature. Zapiski State Hydrologic Institute (Zapiski Gos. Gidrolog. In-ta.). **13:** 95–138. In Russian.

SHOSTAKOVICH, V. B. 1936. Geschichtete Bodenablagerungen der Seen als Klima-Ann. Met. Z. **53:** 176–182.

SHOSTAKOVICH, V. B. 1944. An experiment on geochronological analysis of mud deposits of Malinovoie Lake in connection with the uplift of the shore of the White Sea. Bull. All-Union Geographical Society, U.S.S.R. (Georgraficheskoe obschestvo SSSR, Izvetsia). **76:** 203–206. In Russian.

SCHULMAN, E. 1945. Tree ring hydrology of the Colorado River Basin. Univ. Arizona Bull. 16, no. 4. Lab. of Tree Ring Research Bull. **2:** 35.

SCHULMAN, E. 1956. Dendroclimatic Changes in Semiarid America. Univ. Ariz. Press. Tucson, Ariz. 142p.

SEIBOLD, E. & R. WIEGERT. 1960. Untersuchungen des zeitlichen Ablaufs der Sedimentation im Malo Jezero (Mljet, Adria) auf Periodizitäten. Z. Geophys. **26:** 87–103.

STETSON, H. C. 1933. Scientific results of the "Nautilus" expedition, 1931; part 5, the bottom deposits. : 17–37. Cambridge, Mass.

STETSON, H. T. 1947. Sunspots in Action. Ronald Press. New York, N. Y.

SUGAWARA, K. 1934. Liesegang's stratification developed in the diatomaceous gyttja from Lake Haruna, and problems related to it. Bull. Chem. Soc. Japan. **9:** 402–409.

UDDEN, J. A. 1924. Laminated anhydrite in Texas. Bull. Geol. Soc. Am. **35:** 347–354.

UDDEN, J. A. 1928. Study of the laminated structure of certain drill cores obtained from Permian rocks of Texas. Carnegie Inst. Wash. Year Book. **27:** 363.

VAIL, O. E. 1917. Lithologic evidence of climatic pulsations. Science, (n.s.). **46:** 90–93.

WARD, F. W., JR. 1957. Power spectra of astrogeophysical and meteorological time series. Air Force Cambridge Research Center. Cambridge, Mass. 144p. Unpublished.

WELTON, M. 1944. Pollenanalytische, stratigraphische, und geochronologische Untersuchungen aus dem Faulenseemoos bei Spiez. Geobotanisches Institut Rübel. Veröffentlichungen. **21:** 201.

WILLET, H. C. 1953. Atmospheric and oceanic circulation as factors in glacial-interglacial changes of climate *In* Climatic Change, Harlow Shapley, Ed. 318 p. Harvard Univ. Press. Cambridge, Mass.

UPPER CRETACEOUS AND TERTIARY CLIMATIC PERIODICITIES AND THEIR CAUSES

Th. van der Hammen
Geologisch Instituut, Leiden, The Netherlands

The Facts and Their Interpretation

It has long been known that climatic variations occurred in the Tertiary and older periods. However, these recognized changes were of a general and long-lasting character as, for instance, the gradual cooling of the climate both on the Northern and the Southern Hemisphere since the Middle Tertiary. Pollen analysis has opened the possibility for the first time of also making detailed investigations on eventual climatic fluctuations of a minor order. One can study now the changes of the flora and vegetation at relatively short time intervals if the sediments are favorable. Very few continuous diagrams with sufficiently close sample distance and extending over appreciable lapses of time have been published up until now; we are only at the beginning of a very promising epoch of development of these possibilities of pollen analysis.

Let us see what our own investigations have brought to light, compare them with those of others, and see to what conclusions they are leading.

A composed but continuous diagram from Colombia (South America), comprising the Uppermost Cretaceous and a greater part of the Tertiary was published a few years ago (van der Hammen, 1957). In this diagram species of palms, other angiosperms, and ferns are divided into eight groups. It is not pertinent here to enter into the details of the diagram and diagram construction; therefore we shall only put forward the main facts. Several curves of groups of the diagram show maxima and minima at rather regular intervals but, especially, a group of certain palms (the Monocolpites medius group) show very well marked maxima (FIGURE 1), often combined with maxima of one or two other groups. Upon comparing diagrams of sections from different parts of Colombia (as far as 1500 km. apart), and representing different facies, it became clear that the above-mentioned maxima were constant and synchronous and therefore excellent correlation horizons. As facies differences, transgressions, and other local influences could be excluded, it seemed most probable that the maxima were caused by climatic changes.

The Monocolpites medius group manifested some general fluctuations besides the short maxima, showing higher values in the Paleocene and the Middle Eocene. The curve, taking away the short maxima, was remarkably similar to Dorf's temperature curve for the western United States, based on plant macrofossils (FIGURE 2). High M. medius values would thus correspond to a lower average annual temperature. Several other facts support this conclusion (among them, the important lowering of temperature at the base of the Paleocene indicated by Umbgrove, 1942, corresponding to the most important maximum of M. medius in the whole diagram), so that there seems to be no doubt about its correctness. A rapid look at the diagram will convince us that the M. medius maxima are not irregularly placed and high. In Maastrichtian, Paleocene, and Lower Eocene, for instance, the distances between

the tops are remarkably constant; then the distances become shorter and afterward longer again but, nevertheless, rather regular. The regular distances are to be observed in sediments of times with a quiet sedimentation, and the shorter ones in sediments of tectonically important times (the thin Middle Eocene).

Moreover we may observe that every third top is always higher than the others. This fact, especially clear in the lower part of the diagram, together with the regular distances, seems to indicate a periodic phenomenon. The importance of the maxima becomes especially clear, as it could be established that the high maxima indicate the very base of international units like the Paleocene, Lower Eocene, Middle Eocene, Upper Eocene, Lower Oligocene, and the rest, and that an important change of flora can be observed exactly at the same time (it is especially clear at the bases of the Paleocene, the Eocene, and the Oligocene, where many new types appear for the first time, and the whole floral picture changes).

All the above-mentioned facts seem to point to periodic decreases of temperature of more than ordinary importance, because they are related with international—that is, world-wide—stratigraphical time units. If we suppose that the periods are of the same or approximately the same length of time (which is plausible for more than one reason), then it is easy to calculate their approximate duration. If we take the base of the Tertiary at 60 million years and the base of the Miocene at 18 million years, then we have 21 intervals in a 42-million-year interval. Every interval between 2 temperature minima would then be approximately 2 million years, and the major intervals between the higher maxima 6 million years. Although these figures may not be exact and, even if the intervals were a little more irregular in some times than we supposed, these figures should give at least a good approximation.

A periodicity of temperature decreases at more or less regular intervals of 6 and 2 million years, corresponding to international time units; that seems to be the conclusion we can draw from the diagram.

Although the diagram does not include deposits younger than the Lower Miocene, there are several clear indications that the same rhythm continued during the whole Miocene and the Pliocene. A very important decrease of temperature is shown in the diagram at the base of the Paleocene (the highest M. medius maximum of the whole diagram). Approximately 60 million years later, during the Pleistocene, an even more important lowering of temperature causing the Quaternary ice ages took place. Umbgrove (1942) mentioned cold periods at the beginning of the Paleocene and some 60 million years earlier at the beginning of the Cretaceous. It may be therefore that the strong temperature decrease we established at the base of the Tertiary belongs to a still longer periodic phenomenon.

After the foregoing one may ask if these periods are world-wide or only local. We may answer this question in two ways: (1) by simple theoretic considerations, and (2) by looking for similar phenomena in diagrams from other parts of the world. First, the more important temperature minima lie exactly at the base of international stratigraphic time units. From this it will be clear that there has to be some causal relation between these minima and the

rhythm of life, on which the international units are based. Moreover this is clearly proved by the diagram itself, showing the exact relation of the temperature minima with the disappearance of species from the former period and first appearance of new species. From such a continuous and exact connection with world-wide time limits we cannot draw but one conclusion: the world-wideness of the phenomenon itself.

Do we really find the same phenomenon in other parts of the world? Unfortunately there has not been published much data that can serve for comparison. The continuous long-time diagram is still rare in literature. However, very good diagrams from the Pliocene and Lower Pleistocene were published by Zagwijn (1960). The climatic interpretation is difficult, but a few important facts can be established with certainty.

(1) There is a very clear cold phase at the base of the Quaternary. In

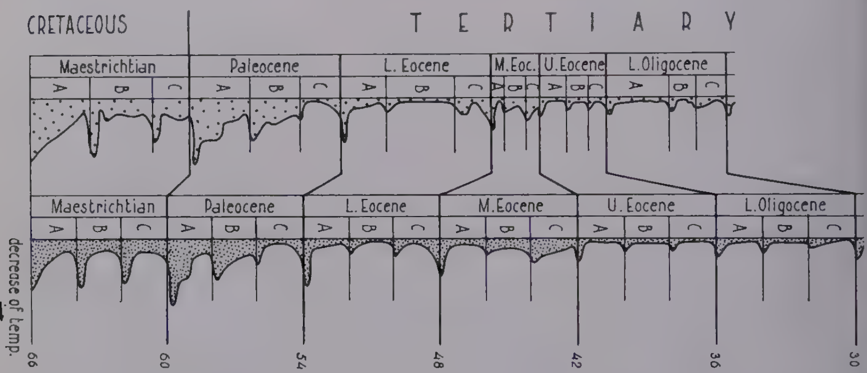


FIGURE 1. Curve of the Monocolpites medius group, directly taken from the diagram temperature curve (*right*). The dotted line (for the Miocene and Pliocene) is deduced from

FIGURE 3 we have redrawn and somewhat simplified this part of Zagwijn's diagram, comparing it with the cold phase at the base of the Tertiary. Very interesting is the sharp fall in both cases of respective "Tertiary" and "Cretaceous" elements.

(2) The Pliocene was subdivided into three more or less equivalent parts: from young to old the Reuverian, the Brunssumian, and the Susterian. There is a clear short cold period at the base of the Reuverian (Tertiary elements very low, Pinus high, for example). The same tendency seems to be indicated in the diagram at the limit of the Susterian and the Brunssumian, although less pronounced. No data are available from the base of the Susterian, as these are sand layers, but a cold phase must be supposed here, according to Zagwijn, to explain the disappearance of certain thermophile Miocene elements.

(3) Also at the base of the Upper Miocene a clear cooling of the climate seems present; the Miocene elements fall steeply, while the "Pleistocene elements" show a rapid rise.

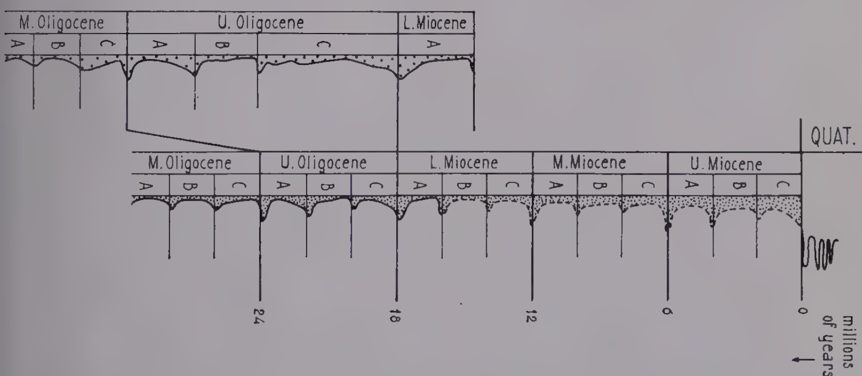
Although Zagwijn is still working on a more definite and detailed climatic

interpretation of his excellent diagrams, we think that the above-mentioned facts already allow us to establish the presence in the European Miocene and Pliocene (including also the base of the Quaternary) of lowerings of temperature at intervals, which are of the same nature as those established in the Uppermost Cretaceous and the Tertiary of Colombia.

As yet unpublished investigations on Paleocene material from British Guiana disclose that similar fluctuations of the M. medius group to those established in Colombia occur.

Other unpublished palynological studies by P. M. J. van Hoeken-Klinkenberg on material from the Uppermost Cretaceous of Nigeria (Africa) also give very important data. Exactly there where the Paleocene should begin, the M. medius group shows a well-marked and sudden rise, just as in the Colombian diagram on the same place.

There seems no doubt that the future will bring us many more data and confirmation on the world-wide and synchronous nature of the fluctuations. We



(*left*). The same curve redrawn on a time scale and somewhat simplified, as a relative other diagrams and different facts, and does not pretend to be exact.

are only at the beginning of the development of pollen analysis in this direction, but perhaps in the near future we shall be able to make direct and exact intercontinental correlations by means of periodic temperature decreases; the first steps have been made already.

We may ask if the phenomenon occurs in formations older than the Uppermost Cretaceous. There seems no reason why it should not, as it could be established to have occurred during almost 70 million years. Certainly there are indications that something similar may be present in the rest of the Cretaceous. Bürgl (1959) recognized 10 cycles of approximately equal thickness of sediments in the relatively quiet geosynclinal facies of the Cretaceous in the Eastern Cordillera of Colombia, on a paleontological basis corresponding with the international time units: Maastrichtian, Senonian (= Campanian + Santonian + Coniacian), Turonian, Cenomanian, Albian, Aptian, Barremian, Hauterivian, Valanginian, and Berriasian. This author assigned an approximate duration of 6 million years to these 10 phases, based on a 60-million-year duration of the Cretaceous.

In this respect we may mention again the cold phase indicated by Umbgrove

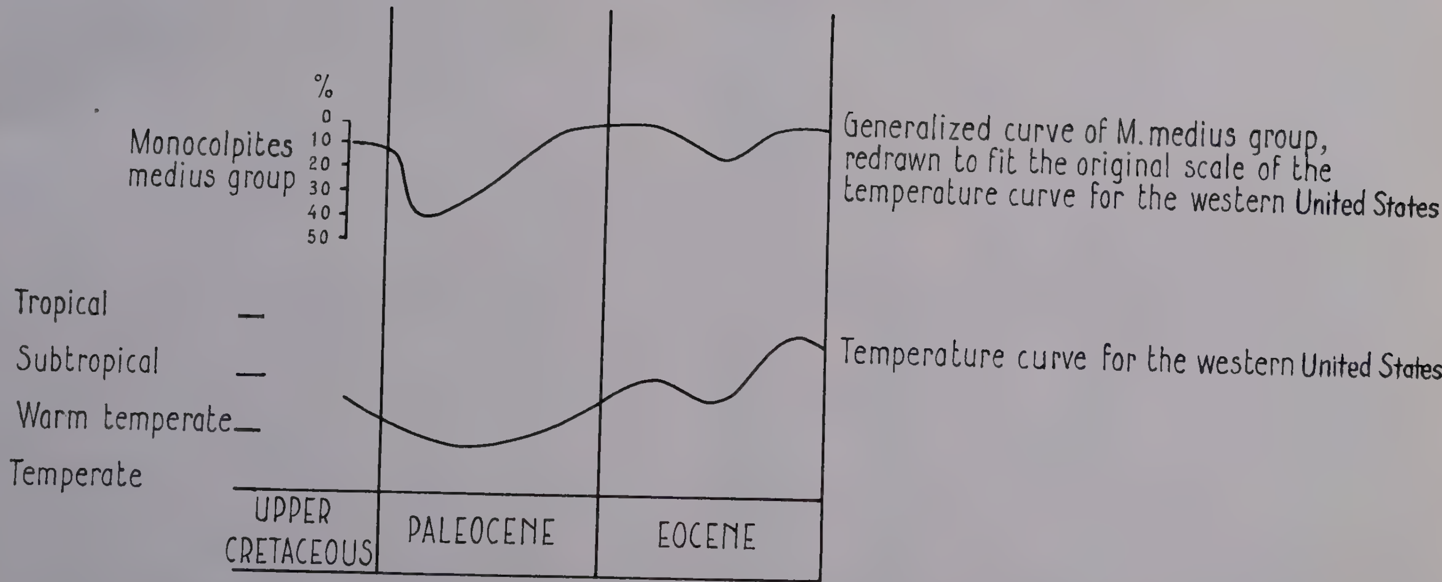


FIGURE 2. Temperature curve for the western United States, based on plant macrofossils, compared with a curve based on the frequency of the Monocolpites medius group, based on a pollen diagram from Colombia.

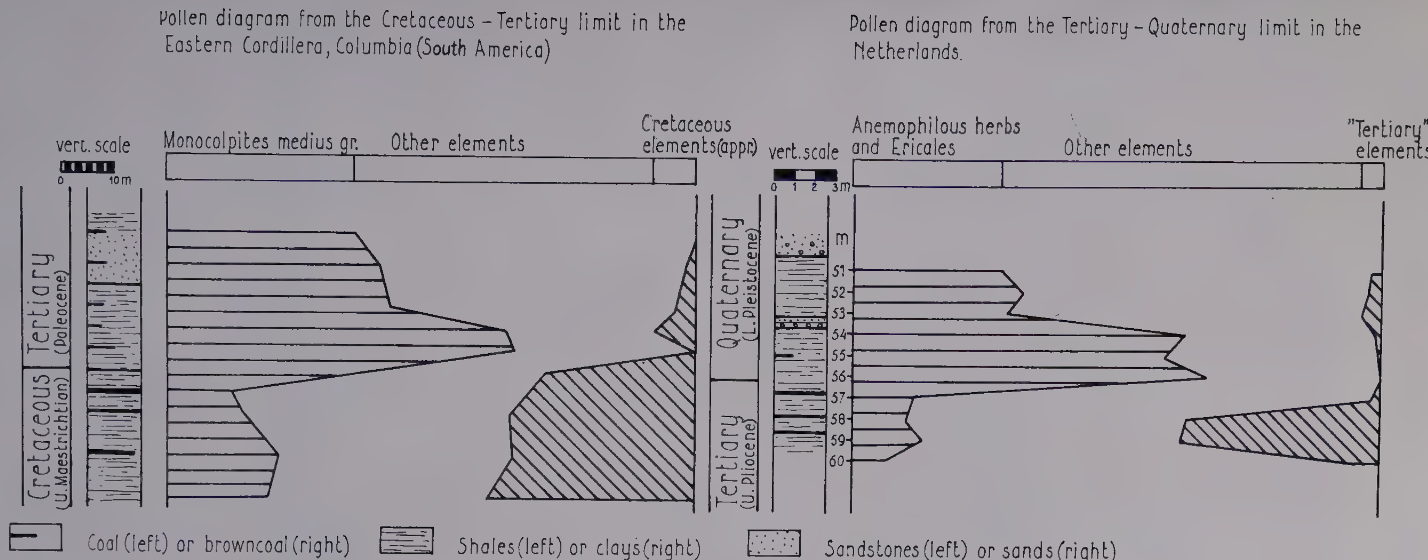


FIGURE 3. Temperature decrease at the beginning of the Tertiary (*left*) and the Quaternary (*right*), as shown by pollen diagrams. The group of pollen represented by horizontal lines (left part of each of the diagrams) indicates temperature changes: a higher percentage corresponds to a lower temperature. Right graph adapted from the diagram "Meinweg" in Zagwijn (1960).

(1942) at the beginning of the Cretaceous, similar to the one at the beginning of the Tertiary. Similar cold phases are indicated by that author at the beginning of several periods of the same order. Investigations have been started now in Leiden, The Netherlands, in order to determine whether this cold phase at the beginning of the Cretaceous can be recognized by means of pollen analysis.

Other indications may be found in the excellent statistic palynological investigations on the Carboniferous by Dybova and Jachowicz (1957). Peculiar sudden changes of pollen flora may be observed in their diagrams, repeated at intervals. R. H. Wagner (personal information), on the basis of his studies of carboniferous stratigraphy and macroflora successions, also thinks that marked repeated temperature changes occur throughout the Upper Carboniferous.

All these are of course only indications, but sufficiently important to keep in mind for future investigations the possibility that climatic periodicities of approximately 60, 6, and 2 million years might have exercised their influence throughout at least a greater part of geological history. These periodicities would then find their natural place in decreasing order of magnitude between the 200- or 250-million-year periodicity of the great ice ages at one side, and the glacial-interglacial and still smaller climatic rhythms at the other.*

Secondary Effects

Pleistocene temperature fluctuations produced falls and rises of the sea level, caused by the increase and decrease of icecaps on the continents. Tertiary and older temperature fluctuations must have had a similar effect. Even if under normal conditions there were no polar icecaps, the colder phases might have produced them, even if they were small. The influence of such appearing and entirely disappearing icecaps on the temperature of the ocean water (even in equatorial regions) and its animal life must then have been considerable. There are many indications that these theoretical deductions are right.

The fact that the ocean water suffered decreases of temperature at intervals may be observed in the oxygen isotopic temperature curves for Lower-Middle Miocene and Middle Oligocene deep-sea cores. The maximal temperature differences are 3° C. (naturally a general temperature decline affecting the whole earth might also directly influence the surface ocean water).

It is a well-known fact that the end of an international time unit is often characterized by a regression of the sea, and the beginning of the next one by a successive transgression. As we have seen that cold phases occur exactly at the limits of those units, it seems logical to correlate the cooling of the climate with the regression, and the successive warming up with the transgression. This of course neither excludes eventual tectonic causes nor epirogenetic movements, at the same time or separately. Nevertheless it is another important indication that the supposed relation must be right, that in the Miocene sequence of the Caribbean coast in Colombia a series of regressions and transgressions could be established, with a rhythm that resembles very

* Eventual periodic temperature changes with an interval smaller than 2 million years could not be established in our Tertiary-Cretaceous diagram, because the sample distance was too great for the purpose. They might very well be found by more detailed investigations.

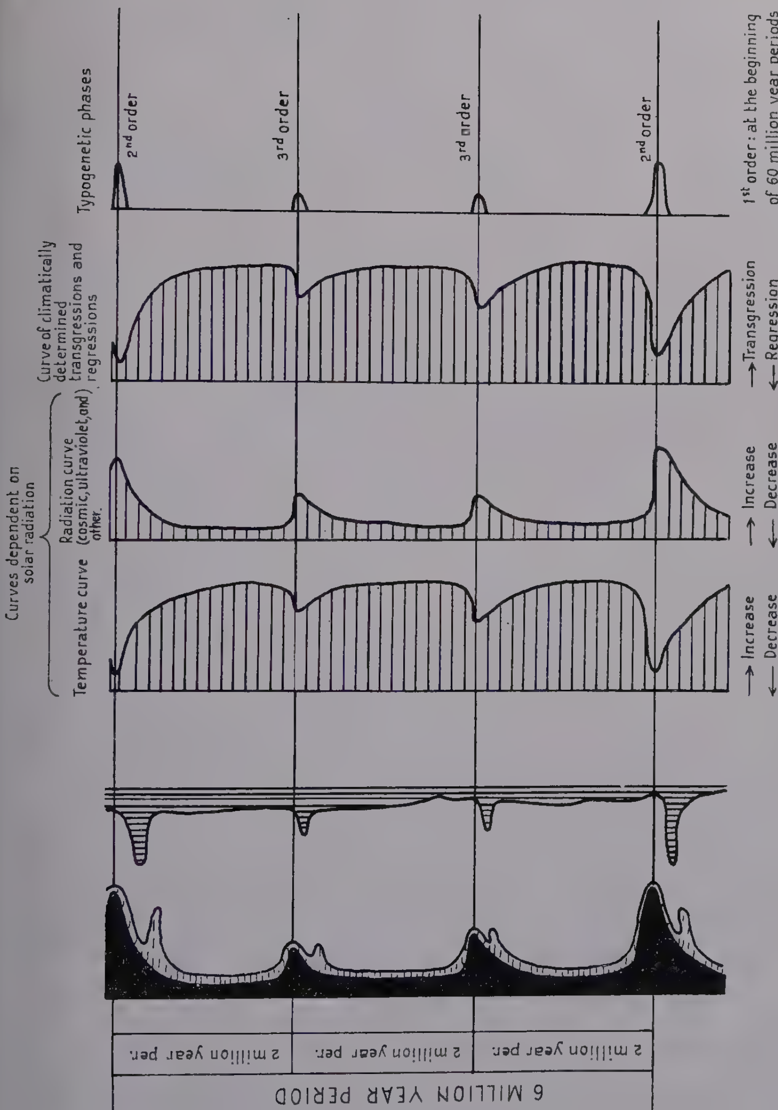


FIGURE 4. Supposed correlation of the quantitative vegetation changes, fluctuations of temperature and radiation, climatically determined transgressions and regressions, and typogenesis.

much that of the temperature decreases, from old to young successively, an important transgression followed by two less important ones and again an important transgression followed by two less important ones. The sediments between two transgressions are in all cases remarkably little differing in thickness from each other, and represent according to Bürgl's interpretation (Bürgl *et al.*, 1955) international time units (Aquitanian, Lower Burdigalian, Upper Burdigalian, Lower Helvetian, and others).

If the above statements are true, and it seems very likely, then we could eventually expect the same relation between movements of the sea-level and temperature changes in many older formations, especially for instance in the Permo-Carboniferous, when even glacial and interglacial conditions occurred.

Causes

If we like to explain the facts and find out the causes, we should first consider what the theory should make clear.

There are two principal facts to be explained: (1) the probably periodic decreases of temperature affecting apparently the whole earth; and (2) the clear and exact relation of these decreases of temperature with the limit of international stratigraphic time units or, in other words, with the "clock of life," depending on the evolution process (typogenesis).

After considering both internal and external causes carefully, I find it impossible or extremely difficult to imagine that causes from inside the earth could have originated these rhythmic effects on climate and life at the same time in such a profound way.

Again, just as in the explanation of the glacials and interglacials of the Quaternary, fluctuations of solar radiation seem able to give a really satisfactory explanation in our case; and even more probable, because it seems impossible to explain the contemporaneous temperature decreases and "typogenetic" phases of evolution by one and the same cause in another way as by solar cycles—cycles with periods of lower heat radiation and higher emission of one or more mutation-creating radiations (cosmic radiation and ultra violet radiation, for example; see FIGURE 4). Long range solar cycles seem the only possible and logical phenomenon that can explain all the facts, according to our present knowledge. However, these cycles would be so long (2 and 6 million years and, eventually, 60 million years) that it will be impossible to prove their existence by direct observations. Nevertheless it seems that the facts discussed above oblige us to accept their existence.

References

BÜRGL, H. 1959. Bosquejo de la geologia de Colombia y Suramerica. Servicio Geologico Nacional, Internal Rept. Unpublished.

BÜRGL, H., M. BARRIOS & A. M. ROSTROM. 1955. Micropaleontologia y estratigrafia de la seccion de Arroyo Saco, Departamento del Atlantico. Bogotá. Bol. Geol. **3**(1).

VAN DER HAMMEN, T. 1957. Climatic periodicity and evolution of South American Maastrichtian and Tertiary floras. Bogotá. Bol. Geol. **5**(2): 49–91.

UMBGROVE, J. H. F. 1942. The Pulse of the Earth. Martinus Nijhoff. The Hague, Netherlands.

ZAGWIJN, W. H. 1960. Aspects of the Pliocene and Early Pleistocene vegetation in the Netherlands. Mededelingen van de Geologische Stichting. **C-iii-1**(5): 1–78. Haarlem, The Netherlands.

DYBOVA, S. & A. JACHOWICZ. 1957. Microspores of the Upper Silesian coal measures. Warszawa, Inst. Geol. **23**.

PALEOCLIMATIC IMPLICATIONS OF PLEISTOCENE STRATIGRAPHY IN THE MEDITERRANEAN AREA*

Karl W. Butzer

Department of Geography, University of Wisconsin, Madison, Wis.

Introduction

Studies in climatic variation may consider both the *causes* of indicated climatic changes and the *patterns* of such changes. Often, however, the patterns are neglected, and far-reaching conclusions are drawn on the basis of disintegrated and sporadic evidence. This is of course inevitable in a field of research embracing several distinct disciplines. It is nevertheless necessary to elaborate on the distribution of climatic anomalies both in time and space if a reliable perspective of general circulation mechanisms is to be obtained. Analyses of recent climatic fluctuations registered within the instrumental record have made considerable progress in this channel lately. In the Pleistocene, however, almost all interpretative study has made use of hopelessly inadequate schemata of glacial, interglacial, pluvial, interpluvial. The major circulation changes of the Quaternary were anything but a simple expansion or contraction of climatic belts. The differential variation of the various climatic elements at any one locality has proved to be astoundingly complex whenever detailed evidence became available.

Consequently the following materials will be devoted to a brief summary of the patterns of climatic evolution in the zone of overlap between the subtropical cells of high pressure and the circumpolar westerlies in the general area of the Mediterranean Basin. Outlining the earth science evidence involved, several climatic patterns, related to specific types of "glacial" and "interglacial" circulation of the atmosphere, will be discussed.

The provisional results presented here were obtained during various field sessions in the Near East and Mediterranean area since 1956, and much of the material has not yet received publication. On account of limited space, discussion and documentation have necessarily been reduced to a minimum.

"Tropical" Interglacial Climates

Perhaps the most fundamental realization in understanding Quaternary climates is that present Postglacial climate is not the only and perhaps not the dominant interglacial climatic pattern in this area.

Stratigraphically associated with the Tyrrhenian I or Holstein Transgressions (Butzer and Cuerda, 1961*a* and *b*) or overlying outwash deposits or moraines of Mindel age (Fränzle, 1959) is a mature soil of striking *Rotlehm* characteristics common to the Mediterranean, to much of temperate Europe and, in a limited way, to the northern Sahara (Butzer, 1959). The physical and chemical properties of this soil are so individual that they provide a Leithorizon, for example, the north Italian *ferretto*. On limestone this *Rotlehm* represents a *Terra rossa* whose (B)-horizon may attain 5 m. depth in moister areas, 50 to

* Parts of the contents of this paper were given at colloquia at the Geographical and Geological Institutes of the University of Cologne, Cologne, Germany (January 11, 1961) and the Geographical Institute, Bonn, West Germany (January 16, 1961).

100 cm. in drier parts. On acidic bedrock a true *Rotlehm* plastosol is present. Unfortunately the mechanics of this soil development are but imperfectly understood. Reifenberg (1947) was able to show that the sesquioxides are carried in solution by silicic acid and washed downwards during wet periods, and carried upwards by capillary action in succeeding dry phases, and then precipitated. Although no modern analogies are known, the intensity of the chemical weathering inherent to the first part of the process is presumably possible only under warm and humid conditions. The second phase certainly requires a periodically dry climate. Although soil studies are lacking there, such conditions are theoretically met in the Hyrcanian rain forest on the southern shores of the Caspian Sea, an area with a subtropical vegetation enjoying a high precipitation, with one distinctly arid month annually.

Much remains to be clarified in the stratigraphy of these *Rotlehms* before their last major development can be unequivocally set in the "Great" (Mindel/Riss) Interglacial. Corroborative evidence is available in the monograph of J. H. Durand (1959) on Algeria, and the writer has observed such soils developed specifically on the "high" (Mindel) pluvial terraces of rivers in central Catalonia (unpublished), and on the (early Holstein) Handborough terrace of the Thames at Oxford (Sandford, 1929). *Terra rossa* development of Eem (Riss/Würm)-Tyrrhenian II date has been verified on Mallorca (Klinge and Mella, 1957; Butzer and Cuerda, 1961*a* and *b*) but it is of limited importance: the climax soil tends towards a *Braunlehm.*

The similarity of these soils from North Africa to Central Europe suggests a considerably more uniform zonation than today, a moist, warm climate with a short but appreciable dry season. Paleobotanical evidence bears this out: Mediterranean species in the Inn valley of Austria, an extension of numerous thermophile species into Scandinavia and the Union of Soviet Socialist Republics, and above all the pollen diagrams, in which *Pinus* plays an unproportionately large role compared with Eem or Postglacial profiles (cf. Woldstedt, 1958). Deep-sea core measurements suggest the tropics were 1° C. warmer than today, the paleobotanical evidence in Europe suggests 2.5 to 3° C. warmer. Similarly the glacial-eustatic rise in sea level to altitudes well above those of the present (although obscured by the progressive lowering of Quaternary sea levels) seems to suggest a slightly greater melting of the residual icecaps. The meteorologic inference would be a reduction of the temperature gradient between pole and equator with attendant atmospheric corollaries. Consequently this paleoclimatic type bears much similarity with a nonglacial climate. It may be recalled that preglacial climates were responsible for intensive kaolin weathering in temperate Europe, and related phenomena exist in the Mediterranean and even the Sahara. These are nonglacial pluvials of the Tertiary (Butzer, 1957*b*).

"Subtropical" Interglacial Climates

Intercalated with the Tyrrhenian II or Eem Transgression of distinctly thermophile fauna (Butzer and Cuerda, 1961*a* and *b*), overlying outwash or moraines of Riss age (Fränzle, 1959), and on "middle" (Riss) pluvial terraces in Catalonia (Butzer, unpublished) are deposits of soils developed *in situ*

suggesting a modified climate of the type mentioned above. The climax soil development was a *Braunlehm* plastosol or *Terra fusca* whose genesis is as uncertain as that of the *Rotlehms*. In the southern Mediterranean an added feature is the occurrence of tufaceous calcareous crusts (*croûtes zonaires*) due to precipitation in lime-charged surface waters. This phenomenon has been studied in detail by Durand (1959), as well as by myself in Mallorca (unpublished). The most ready hypothesis is that of sheetflooding under lightly stocked forest, dissolution of carbonates by humic acids, and eventually sedimentation of fine laminae of calcite with release of CO_2 by the waters under warm conditions. Frequent embedding of originally dehydrated crumbs of *Terra fusca* in these crusts suggests storm rains after drought periods, with erosion of contemporary soils around cracks due to contraction of parched clays. The evidence of this limited erosion and of a diminished intensity of pedogenesis suggests an intensification of the dry season, compared with the palaeoclimate discussed previously, possibly with a moderate reduction in over-all precipitation.

On Mallorca moderate soil developments and such sedimentary crusts are further associated with the interstadials of the Last Glaciation and with the maximum of the Flandrian Transgression* (Butzer and Cuerda, 1961*b*), and in Italy (Fränzle, 1960) and Central Europe (Remy, 1960) sols lessivés were also typical for Würm interstadials. This suggests a similarity of soil development intensity over wide areas during both "full" interglacials with reduced latitudinal temperature gradients, as well as during intense interstadial climates with the inevitable exaggerated temperature gradients (due to greater extension of the glaciers). Obviously there is much still to be learned here in matters of stratigraphy, interpretation, and paleoclimatic synthesis. The existence of such a little-understood, nonglacial climate of moderate intensity is however unquestionable.

"Temperate" Interglacial Climates

If we turn to the expression of recent Postglacial climate on soils and geomorphology a considerable reduction in moisture must be assumed for the Mediterranean area in comparison to the previous interglacial paleoclimates. Apart from saline anhydration there is almost no weathering in the Sahara today (Butzer, 1959; Meckelein, 1959); only xerorendzinas or brown earths develop in the Mediterranean Basin (Klinge and Mella, 1957; Durand, 1959, Butzer and Cuerda, 1961*b*; Fränzle, 1959), and even in Central Europe the Postglacial soils may be less well developed than those of the "Göttweig" Interstadial. On Mallorca the ancient plastic soils have suffered a migration of colloidal clays since the Subboreal (Butzer and Cuerda, 1961*b*) and throughout the southern Mediterranean, morphogenesis is almost inactive wherever man has not induced accelerated erosion.

This morphostatic character of the prevailing climate in the subtropics is such as to be next to unrecognizable in the fossil record. At best it would be registered as a negative, arid phase conveniently designated as an "interpluvial." Beyond doubt large parts of the Pleistocene interglacials were

* Atlantic phase?

similar in tenor, and the Late Glacial of the Würm left analogous geomorphologic traits. This indifferent array of phenomena may in fact have dominated the greater part of the Quaternary, with the aberrant paleoclimates outlined above and below playing a more spectacular but less persevering role.

Early Glacial or "Pluvial" Climates

In contrast to the foregoing essentially morphostatic paleoclimates of variable weathering intensity, the two climatic patterns of glacial type were responsible for active morphogenesis. Due to active erosion and sedimentation, soil development was physically retarded despite intense weathering. Consequently the fossil record is quite different with conspicuous geological deposits and geomorphological sculpturing.

Throughout the Mediterranean and North Africa—in areas not affected by periglacial climates—the Early Würm glacial left deposits of moderately to well-rolled gravels in the form of fluviatile terraces or alluvial fans, as well as sheets of colluvial silts (*limons rouges*). Morphometric analyses of the gravels rule out a torrential-type, semiarid aggradation (2r/L index maxima at 300 to 350 for fluviatile gravels in Mallorca and Catalonia). Steep downstream gradients in coastal areas suggest a dropping base-level (regressive oceans), a fact that is confirmed by intercalation with and semiconformable overlying of final Tyrrhenian beach deposits (Butzer and Cuerda, 1961*b*). As these gravels occur ubiquitously in the Mediterranean, Near East (Butzer, 1958), and the northern Sahara (Butzer, 1959), also on river systems with interfluves below 300 to 500 m., they cannot be attributed to cold climate or periglacial aggradation. For that matter cryoclastic pebbles or vestiges of solifluction are absent. These then are true *pluvial* gravels.

Related deposits include the colluvial silts or *limons rouges* (cf. Durand, 1959), that contain a large aeolian component in their granulometric spectra (Butzer, in preparation). This is generally due to the transport of older, weathered aeolian materials. The silts often contain angular pebbles and are very frequently interrupted by *croûtes zonaires* indicating clear sedimentation with a temporary decrease in mechanical weathering or erosion. Such *limons* are laterally conformable with true fluviatile deposits in valleys and can frequently be observed to extend well below modern sea level. The general character of these gravels and silts permits an unequivocal association with the Early Glacial of at least the Würm phase.

Corollary processes of similar date are the characteristic stalagmitic horizons of the Mediterranean caves in limestone bedrock (Butzer, 1957*a*), which can best be explained by a considerable percolation of waters through fissures and clefts in the overhanging bedrock, implying a moist climate. This "pluvial" climate need not necessarily have been as moist as that of the "tropical" interglacials, however. The dominating difference is that of mechanical preparation of loose materials (increased frost or thermoclastic weathering, or direct chemicophysical attack by sheetfloods) and greater transport capacities with accelerated runoff. This suggests a more temperate climate with periodic, often semitorrential rainfall.

Chronologically this cooler, periodically-moist phase corresponds with the cold, humid "*Fliesserdezeit*" or major solifluction phase characteristic of the

Early Glacial in central Europe (Büdel, 1950) and the Soviet Union (Frenzel, 1959).

Full Glacial Climates

During the maximum lowering of planetary temperatures the deposition of loess or other aeolian deposits ("*Lösszeit*") in periglacial areas was paralleled by the deposition of regressional aeolianites in the subtropics. The latter is a littoral phenomenon of only qualified paleoclimatic value, as root casts, root drip, and petrified roots are commonly found in association with land snails in such beds. The contemporary continental deposits are more indicative in the Mediterranean area however. Thus for example the Early Glacial gravels go over conformably (in the vertical sense) into finer, poorly stratified and increasingly angular beds, which in their turn are conformable (laterally) with fans of colluvial silts due to intensive areal erosion. Such deposits contain certain quantities of gravel that has been mechanically smashed after rolling. It would be an exaggeration to designate these as "cryoclastic," but the co-agency of cold climate weathering—even in deposits of North Africa (Hey, 1961)—is undeniable. The necessary erosion and transporting ability on the

TABLE 1
WIND *DEVIATIONS* OF STORM WINDS ON THE MALLORCAN LITTORAL

	Southwest coast	Southeast coast
WIa	30° W	9° E
WIb	38° W	27° E
WII (early)	50° W	25° E
WII (later)	(no record)	48° E

other hand requires considerable moisture, although it was somewhat drier than during the previous period, judging by the angular nature of the gravels and, above all, by the *éboulis secs* or moderately fine, uncorroded thermoclastic debris of the Mediterranean caves (Butzer, 1957*a*): a distinctly fossil feature.

The Full Glacial climate induced a less thermophile fauna, chiefly temperate woodland species, with large numbers of rodents, to frequent the Mediterranean areas. The Mediterranean waters were considerably cooler judging by the recent discovery of *Cyprina islandica* in Würm deposits off the Spanish coast, and by the deep-sea core measurements of Emiliani (1955), which indicate a lowering of mean August temperatures by 5° C.

An interesting paleoclimatic sidelight of the regressional dunes are the differences and changes in wind direction they indicate. There are three Würm dunes on the littoral of Mallorca, obviously dating from the beginning to the end of the oscillating regression, that is, they are prior to the retreat of the continental glaciers from the outermost end moraines. On this basis these dunes are tentatively called WIa, WIb and WII (Butzer and Cuerda, 1961*b*). Modern storm wind directions are indicated by inclinations and deformation of shrubs and trees, whereas the fossil wind directions can be synthesized with reasonable accuracy on the basis of bedding directions in view of simple forest bedding. Modern dominant storm wind direction of the southeast coast is toward 105° W of mag. N, of the southwest coast 80° E (TABLE 1).

As the isobaths run roughly parallel to the modern coastlines there is but limited topographical effect. These deviations are due to a different system of winter storms: a sharp decline in the number of gales over the Gulf of Valencia (associated with mP outbursts along routes VIa and particularly VIb; cf. Butzer, 1960), and a larger number of deep cyclonic lows travelling east of Gibraltar along route VII (FIGURE 1). The rapid veer indicated by the Würm II dunes suggests a regular deep low about longitude 5° E north of the Algerian coast during the glacial maximum. This all tends to confirm a southerly shift

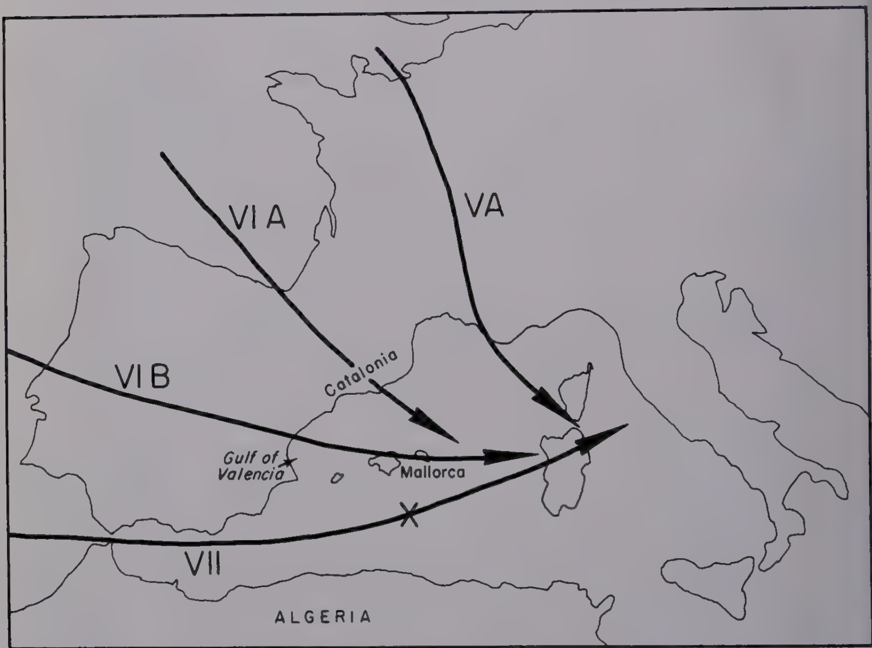


FIGURE 1. Major zones of frontal passage into the western Mediterranean (after Butzer, 1960). At present the frequency ratios of routes Va: VIa: VIb: VII are 2:3:3:1. During the Würm Glacial route VII was more important than VIb, during the Riss the importance of VIa and VIb was greater than at present. X indicates presumed secondary center of cyclonic activity during the Würm Glacial.

of the westerly cyclones. During the various phases of the Riss regression the southeast coast phenomena remain analogous, but the west coast evidence implies a much greater incidence of Valencia gales, even with a more northerly trajectory (route VI6).

Late Glacial Climates

The fossil record in the Mediterranean area indicates that the Late Glacial period was one of limited soil development and inconsiderable morphogenesis, very similar to the "temperate" interglacial climates outlined above. Our reason for mentioning it briefly here is to draw attention to contemporary features in Europe. Poser's (1950) invaluable study of the fossil, Late Glacial

continental dunes of Central Europe enables a reliable reconstruction of the responsible isobaric distribution. This suggests almost the whole range of zonal or mixed circulation types known from contemporary *Grosswetter* research. In view of the reduction of evaporation over the cool glacial age oceans (by more than 25 per cent according to Flohn, 1953) it is easy to understand the prevailing aridity of Mediterranean and Saharan climate (Butzer, 1957*b*, 1958, 1959).

Conclusions and Retrospect

No attempt is made here to evaluate or interpret the earth science material meteorologically (cf. Butzer, 1961). I wish merely to show conclusively *the existence of cool-wet, cold-dry, warm-dry, and warm-moist paleoclimates in the Mediterranean area*, with indications that the anomaly in each case also pertains to temperate Europe and northern Africa and, in some cases, to the Sudanese belt as well. "Mediterranean pluvials" obviously exist and are intimately associated with a particular glacial-type anomaly of the general circulation. Also, south of the Sahara there are lacustrine deposits (*Sahara au Tchad*) apparently of Early Glacial date (cf. Butzer, 1957*b*). Yet the Mediterranean climates have another type of pluvial, a period of intensive chemical weathering responsible for the fossil red soils common also to temperate Europe, North Africa—and in the form of laterites—to the Sudanese savannas. The latter were responsible for the creation of a separate class of "Tropical pluvials" (cf. Balout, 1955), the result of which was a stimulating controversy—based on a misunderstanding—whether tropical and Mediterranean pluvials are contemporary and, if so, whether they are of glacial or interglacial age.

Seen in the perspective of Mediterranean palaeoclimates as outlined here this controversy now resolves itself: "tropical" and "mediterranean" pluvials, designating specific geologic phenomena in the respective areas, are distinct in time but each are common to both the tropics and subtropics. The former are *interglacial* in age, and the latter are *Early Glacial:* without reference to their respective genesis. It should be noted further that the "mediterranean"-Early Glacial pluvials were decidedly less conspicuous in the tropics, which explains why they are often overlooked. Thus the major palaeoclimatic shifts north and south of the Sahara have been, in so far as evidence is available now, *synchronous and not alternative.* The evidence does not support a wandering but an expanding or contracting Sahara.

Again seen within this perspective the "subpluvial" character of the Postglacial Thermal Maximum in the arid zone (Butzer, 1961) does not seem so unusual, while the nature of pre-Pleistocene pluvials in the same area appears to become more obvious. This provides a wide field for purely meteorological research.

References

Balout, L. 1955. Préhistorire de l'Afrique du nord. Arts et Metiers Graphiques. Paris, France.

Büdel, J. 1950. Die Klimaphasen der Würmeiszeit. Naturwiss. **37**.

Butzer, K. W. 1957*a*. Mediterranean pluvials and the general circulation of the Pleistocene. Geografiska Ann. **37:** 48–53.

Butzer, K. W. 1957*b*. The recent climatic fluctuation in lower latitudes and the general circulation of the pleistocene. *Ibid*. **37:** 1–5–113.

Butzer, K. W. 1958. Quaternary Stratigraphy and Climate in the Near East. Dümmler. Bonn, Germany.

Butzer, K. W. 1959. Contributions to the Pleistocene geology of the Nile Valley. Erdkunde. **13:** 46–67.

Butzer, K. W. 1960. Dynamic climatology of large-scale European circulation patterns in the Mediterranean area. Meteor. Rundschau. **13:** 97–105.

Butzer, K. W. 1961. Modifications of climate in the arid zone since the Pliocene. *In* L. D. Stamp, Ed. History of Land Use in the Arid Zone. UNESCO. Paris, France.

Butzer, K. W. & J. Cuerda. 1961*a*. Nota preliminar sobre la estratigrafia y paleontologia del cuaternario marino del Sur y se de la isla de Mallorca. Boll. Soc. Historia Natural de Baleares. **5**.

Butzer, K. W. 1961*b*. Coastal stratigraphy of Mallorca and its implications for pleistocene chronolog in the Mediterranean area. In press.

Durand, J. H. 1959. Les sols rouges et les croûtes en Algérie. Alger. Service des Etudes Scientifiques.

Emiliani, C. 1955. Pleistocene temperature variations in the Mediterranean. Quaternaria **2:** 87–98.

Flohn, H. 1953. Studien zur atmosphärischen Zirkulation in der letzten Eiszeit. Erdkünde. **7:** 226–275.

Franzle, O. 1959. Untersuchungen über Ablagerungen und Böden im eiszeitlichen Gletschergebiet Norditaliens. Erdkunde. **13:** 289–297.

Franzle, O. 1960. Interstadiale Bodenbildungen in oberitaliensschen Würm-Lössen. Eiszeitalter u. Gegenw. **11:** 196–205.

Frenzel, B. 1959. Die Vegetations-und Landschaftszonen Nord-Eurasiens wahrend der letzten Eiszeit und wahrend der postglazialen Warmezeit. I. Abh. Akad. Wiss. Liter. (Mainz). Math. Naturwiss. Kl. Nr. 13.

Hey, R. W. 1961. The Quaternary geology and Palaeolithic archaeology of Libya. Quaternaria. **7**.

Klinge, H. & A. Mella. 1957. Los suelos de las Baleares. Ann. Inst. de edafologia y fisiologia vegetal. **1**.

Meckelein, W. 1959. Forschungen in der Zentralen Sahara. Klimageomorphologie. Westermann. Braunschweig.

Poser, H. 1950. Zur Rekonstruktion der spätglazialen Luftdruckverhältnisse in Mittel- und Westeuropa auf Grund der vorzeitlichen Bennendunen. Erdkünde. **4:** 81–88.

Reifenberg, A. 1947. The Soils of Palestine. Murby. London, England.

Remy, H. 1960. Der Löss am unteren Mittel-und Niederrhein. Eiszeitalter u. Gegenw. **11:** 107–120.

Woldstedt, P. 1958. Das Eiszeitalter. Bd. II. Enke. Stuttgart, Germany.

OUTLINE OF CLIMATIC FLUCTUATION SINCE THE LAST INTERGLACIAL AGE*

Richard Foster Flint
Department of Geology, Yale University, New Haven, Conn.

Friedrich Brandtner
United States Geological Survey, Washington, D.C.

The number and dates of the principal fluctuations of climate since the last major interglacial have been an object of considerable interest during recent years. The important curves published, beginning in 1955, by Emiliani (1958 and titles cited therein), were a pioneer attempt at a graphic representation of climatic changes based on isotopic temperature measurement of sea-floor sediments, with at least partial time calibration. Later several curves based on data from continental stations were constructed by various investigators, and some of them were published. Some are C^{14}-time calibrated, to a first approximation, through a time range of more than 60,000 years.

Using published and unpublished curves, we here compare (FIGURE 1) data from continental stations in Denmark, the Netherlands, Lower Austria, southeastern California, the Great Lakes-St. Lawrence region, and equatorial Colombia. We have used a common scale for time and a scale for temperature that is as nearly uniform as the data permit.

The curves from Denmark and the Netherlands are those published by Andersen, de Vries, and Zagwijn (1960). They are based on pollen analyses, and their far ends, in the vicinity of 70,000 years B.P., are thought by those authors to belong to the end of the Eem Interglacial.

The curve for Lower Austria is based chiefly on a sequence of fossiliferous loess layers and intervening ancient soils. The near end of the curve is based on pollen-analyzed lake sediments and peat layers. Although a large number of exposed sections is involved, the sections agree in showing similar physical and chemical characteristics. Relevant publications include Brandtner (1950, 1954, 1956, 1959) and Fink (1954, 1956). The Fellabrunn soil complex (the Stillfried Complex of Fink, 1954) is equivalent to the time unit known as the Gottweig Interstadial. A critical evaluation of the C^{14} dates obtained by Hessel de Vries from this complex will be published by Brandtner in *Eiszeitalter und Gegenwart* in 1961.

The curve for southeastern California was constructed from data published by Flint and Gale (1958) and Roosma (1958). The segments labeled *Deep Lake* represent sediments deposited in water that was deep, although saturated, most of the time, with carbonate compounds in solution. The segments labeled *Desiccation* represent evaporites. The amplitude of fluctuation of temperature and/or precipitation is not known from the physical evidence. The rectilinear form of the graph reflects abrupt changes in type of sediment, not rate of change of temperature and/or precipitation. If rates of change were known, the graphic slopes would be curves rather than rectilinear in form.

* Parts of this article, which is published by permission of the Geological Survey, Washington, D.C., have appeared elsewhere previously (Flint and Brandtner, 1961).

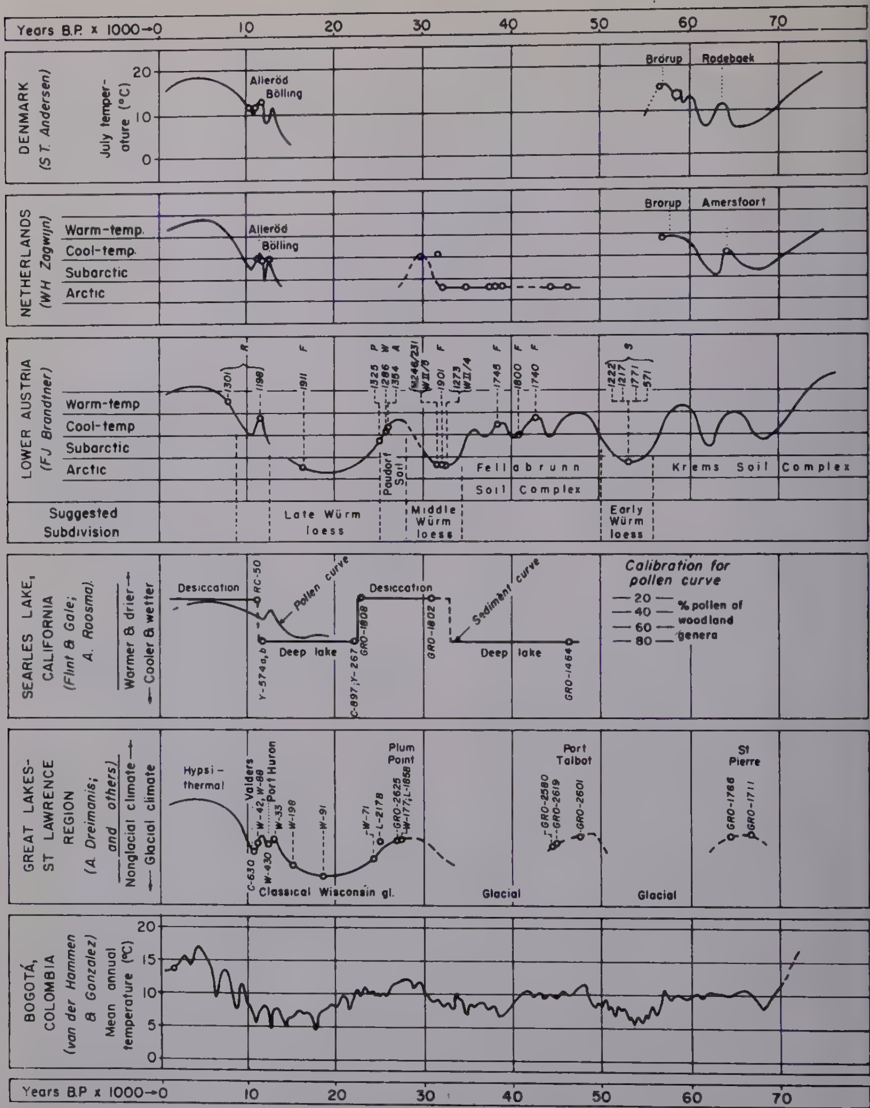


FIGURE 1. Climatic variations in middle north latitudes and at an equatorial station, within the last 70,000 years, inferred from terrestrial data. Circles show positions of C^{14}-dated samples.

Denmark and Netherlands curves (reproduced directly from Anderson, de Vries, and Zagwijn, 1960) are based on pollen data; continuous lines indicate continuous pollen stratigraphy. Temperature calibration differs somewhat between the two curves.

Lower Austria curve, composite for a region, is based on a stratigraphic sequence of loess sheets and soils. The C^{14} date prefixed *H* is from the Heidelberg laboratory; all the others are from the Groningen laboratory. Localities: *F*, Fellabrunn; *P*, Pollau (= Pavlov); *W*, Unter Wisternitz (= Dolní Věstonice); *WII/5* or *4*, Willendorf site II, layer 5 or 4; *S*, Senftenberg; *R*, Roggendorf.

Searles Lake curves are based on physical/chemical characteristics of lake sediments and on pollen, as indicated. Pollen curve is from Roosma (1958).

Great Lakes-St. Lawrence curve is composite, having been compiled from a variety of sources (notably de Vries and Dreimanis, 1960) within a rather wide region. The two pre-classical-Wisconsin glacial intervals are based on glacial drift; the nonglacial segments on peat and gyttja.

Colombia curve (from van der Hammen and Gonzalez, 1960, p. 305) is based on pollen data.

This is evident from the related curve constructed by Roosma from pollen contained in the sediments.

Fluctuations of climate antedating the youngest deep-lake phase and extending back considerably beyond the 70,000-year limit shown in FIGURE 1 are indicated by the stratigraphy (Smith, 1958). Although we accept the fluctuations, we have omitted them from our curve because the available C^{14} dates seem inadequate for control of their positions.

Although we have not included the curve developed from Lake Bonneville sediments by Eardley and Gvosdetsky (1960) because it is not controlled by finite dates, it shows considerable agreement with the Searles Lake curve.

The Great Lakes-St. Lawrence curve has been drawn from various sources. Its far part, antedating 24,000 years B.P., is based on data from de Vries and Dreimanis (1960), Dreimanis (1960, p. 112), and Terasmae (1957). The near part is based chiefly on data assembled by Flint and Rubin (1955) and Flint (1956), from sources cited in those publications. The data do not permit direct temperature calibration, which is relative and suggested only.

The curve from Colombia, the work of van der Hammen and Gonzalez (1960), is based on the pollen record of lake sediments from an altitude of 8500 feet at 4° 30′ N Lat. With the concurrence of van der Hammen, the version of the curve shown in FIGURE 1 was obtained by fitting that part of the curve correlated by van der Hammen and Gonzalez with the Würm Glacial plus the Holocene with the 70,000-year length of the other curves in FIGURE 1. The agreement among major cold and warm intervals is good, as those authors noted (op. cit., p. 307).

As a group, the six curves show good agreement as to their major features, all that can be expected because temperature calibration is not uniform. Probably the Lower Austria curve represents the best over-all documentation, whereas the Colombia curve, although not time calibrated, is the most completely controlled stratigraphically. The similarities among the curves indicate that in several middle-latitude areas of the northern hemisphere and in equatorial South America, climatic fluctuation during the last 70,000 years has been broadly similar, consisting apparently of three principal cold intervals, all younger than the last major interglacial time.

As a group the curves also agree broadly with those developed by Emiliani (1958) for surface-seawater temperature in low and middle latitudes.

References

ANDERSEN, S. T., H. DE VRIES & W. H. ZAGWIJN. 1960. Climatic change and radiocarbon dating in the Weichselian glacial of Denmark and The Netherlands. Geol. Mijnbouw, 39 Jaarg. : 38–42.

BRANDTNER, F. J. 1950. Über die relative Chronologie des jüngeren Pleistozäns Niederösterreichs. Archaeologia Austriaca. **5:** 101–113.

BRANDTNER, F. J. 1954. Jungpleistozäner Löss und fossile Böden in Niederösterreich. Eiszeit. u. Gegenwart. **4/5:** 49–82.

BRANDTNER, F. J. 1956. Löszstratigraphie und paläolithische Kulturabfolge in Niederösterreich und in den angrenzenden Gebieten (Zugleich ein Beitrag zur Frage der Würmgliederung). Eiszeit. u. Gegenwart. **7:** 127–175.

BRANDTNER, F. J. 1959. Die geologisch-stratigraphische Position der Kulturschichten von Willendorf i.d. Wachau, N. Ö. Prähistorisch Kommission der Osterreichischen Akad. der Wiss., Mitt. **8/9:** 173–198.

DE VRIES, H. 1958. Radiocarbon dates for upper Eem and Würm interstadial samples. Eiszeit. u. Gegenwart. **9:** 10–17.

de Vries, H. 1959. Radiocarbon dating of the fossil soils at Ober Fellabrunn. Koninkl. Nederlandse Akad. Wetensch., Proc. Ser. B. **62**(1): 84–91.

de Vries, H. & A. Dreimanis. 1960. Finite radiocarbon dates of the Port Talbot interstadial deposits in southern Ontario. Science. **131:** 1738.

Dreimanis, A. 1960. Pre-classical Wisconsin in the eastern portion of the Great Lakes region, North America. Intern. Geol. Congr., 21st, Copenhagen, 1960 Rept. **4:** 108–119.

Eardley, A. J. & V. Gvosdetsky. 1960. Analysis of Pleistocene core from Great Salt Lake, Utah. Geol. Soc. Am. Bull. **71:** 1323–1344.

Emiliani, C. 1958. Paleotemperature analysis of core 280 and Pleistocene correlations. J. Geol. **66:** 264–275.

Felgenhauer, F., J. Fink & H. de Vries. 1959. Studien zur absoluten und relativen Chronologie der fossilen Böden in Österreich. I Oberfellabrunn. Archaeologia Austriaca. **25:** 35–73.

Fink, J. 1954. Die fossilen Böden im österreichischen Lösz. Quartär. **6:** 85–107.

Fink, J. 1956. Zur Korrelation der Terrassen und Lösse in Österreich. Eiszeit. u. Gegenwart. **7:** 49–77.

Flint, R. F. & F. Brandtner. 1961. Climatic changes since the last interglacial. Am. J. Sci. **259:** 321–328.

Flint, R. F. & M. Rubin. 1955. Radiocarbon dates of pre-Mankato events in eastern and central North America. Science. **121:** 649–658.

Flint, R. F. 1956. New radiocarbon dates and Late-Pleistocene stratigraphy. Am. J. Sci. **254:** 265–287.

Flint, R. F. & W. A. Gale. 1958. Stratigraphy and radiocarbon dates at Searles Lake, California. Am. J. Sci. **256:** 689–714.

Roosma, A. 1958. A climatic record from Searles Lake, California. Science. **128:** 716.

Smith, G. I. 1958. Late Quaternary stratigraphy and climatic significance of Searles Lake, California (Abstr.) Geol. Soc. Am. Bull. **69:** 1706.

Terasmae, J. 1957. Paleobotanical studies of Canadian Pleistocene nonglacial deposits. Science. **126:** 351–352.

van der Hammen, T. & E. Gonzales. 1960. Upper Pleistocene and Holocene climate and vegetation of the "Sabana de Bogotá" (Colombia, South America). Leidse Geol. Mededeel. **25:** 261–315.

CAINOZOIC CLIMATES OF AUSTRALIA

Edmund D. Gill
National Museum of Victoria, Melbourne, Australia

Australia is a continent of about 3,000,000 square miles situated in the Southern Hemisphere partly in the tropic zone and partly in the temperate zone (lat. 10° to 44° S). It is characterized by an arid to semiarid center, with a green border of varying width, including both tropical and temperate rain forests. The Centralian dry area is so large as to be a feature of world importance. Sandridge deserts cover 145,000 sq. mi. in South Australia alone. Along with the deserts of Western Australia and the Northern Territory they constitute one of the most extensive and best developed dune systems in the world (King, 1960). There is thus a great range of climatic conditions (desert to rain forest) in Australia at the present time. During the Cainozoic both colder and hotter, and wetter and drier conditions have been more widespread in Australia. Moreover the present climatic pattern has not obtained for the greater part of that era.

Dry Centralian Area Once Pluvial

It is a paradox that the dry center of Australia abounds in the evidence of former lakes (playas and salinas), including large ones such as Lake Eyre and Lake Callabonna. Most of the "lakes" are permanently dry, while Lake Eyre is usually dry. Only once in recorded history has the bed of the present Lake Eyre been covered with water (Mawson, 1950; Bonython, 1955), but cliffs on its shores consist of late Pleistocene lacustrine sediments (King, 1956), proving a previously more extensive lake. In the floor of Lake Callabonna are the bones of numerous large marsupials. The skeletons of *Diprotodon optatum* (the largest known marsupial, living or fossil); the giant wombat *Phascolonus gigas*; and the giant struthious bird *Genyornis newtoni* have been described from Lake Callabonna, showing that groups of large animals lived in that dry area during the Pleistocene. Similarly, recent investigations have shown the presence of Upper Tertiary faunas in river sediments in Central Australia (Stirton, 1957*a* and *b*). Both the sediments and the fossils provide evidence of comparatively pluvial conditions. Fossil fish, crocodiles and ostracods similarly bear witness to wet conditions in areas now dry.

On the other hand, during the Quaternary there have been times drier and hotter than the present, resulting in extensive desiccation (Butler, 1956, 1959; Gill, 1955, 1960; King, 1956, 1960).

Pedological Evidence of Greater Humidity

Whitehouse (1940, 1948, 1954) described widespread laterites in Queensland that extend to the far southwest of that state toward Lake Eyre, that is, into areas now very dry. Late Cainozoic history is considerably concerned with the breakdown and redistribution of this lateritic material. The extensive ferricretes and silcretes called "duricrust" by Woolnough (1927), and recorded from Western Australian, Northern Territory, and South Australia, are like-

wise lateritic. They imply wet and dry seasons, and prove pluvial conditions for areas now arid. Laterite has also been found in New South Wales, Victoria, and Tasmania. Bauxite widely occurs in Australia (Owen, 1954) indicating deep weathering by copious warm meteoric waters.

Floral and Sedimentational Evidence

The southern beech (*Nothofagus*) is a tree requiring a humid environment. Its typical occurrence is a rain forest on well drained slopes. During the Quaternary pluvial periods it was a little more widespread than now, but during the Tertiary it was common in Australia and had numerous species (Cookson, 1959). Tertiary brown coals and other highly carbonaceous sediments are widespread in the continent. They imply plentiful vegetation and therefore humid conditions. Some of these deposits occur under terrains that are now very dry. Tertiary river beds preserved under basalt flows provide evidence of pluvialty both in their sediments and their fossils (for example, tree trunks and conifer fruits).

Evidences of widespread pluvialty in Australia during the Tertiary and intermittently during the Pleistocene are unmistakable, but there is need for mapping these occurrences in such a way as to show their exact extent for each geological period. This involves the problem of dating these deposits, which is now being gradually done, mainly by palynological methods. The present widespread aridity is a Quaternary intermittent condition, and in this sense is new. The present climatic pattern is atypical of the Cainozoic.

Temperature Changes Inferred from Faunas

Fossiliferous marine sediments of Tertiary age occur from northern Tasmania (ca. 41° S) to the north of Australia. Localities in the present temperate zone contain genera characteristic of the tropical zone, for example lepidocycline foraminifera and echinoderms such as *Phyllacanthus*, *Schizaster*, *Eucidaris*, and *Lovenia*. In southern Victoria, within 1° of latitude, marine fossils of every period of the Cainozoic era are found. They are characterized in mid-Tertiary by tropical forms with a gradual falling off of temperatures both in the lower and upper Tertiary, especially the latter. Thus in the uppermost Miocene (Cheltenhamian Stage) there is a distinct loss of tropical forms, but nevertheless many warmer water forms (including vast numbers of *Lovenia*) are present (Gill, 1957*a*). In the Lower Pliocene (Kalimnan Stage) there is still evidence of warmer waters, but this is not immediately obvious, as it is in the middle Tertiary stages. Foraminifera provide some of the evidence (Parr, 1939), but the obvious giant forms have disappeared. There is a wealth of the gasteropod *Tylospira* and the echinoderm *Arachnoides*, both now limited to waters further north. In the uppermost Pliocene, the fossils indicate slightly warmer waters than now (Gill, 1957*b*), and such evidence is also found in the Pleistocene interglacial deposits (Collins, 1953). As expected, the intervening periods were colder than the present.

The curve of changing Tertiary temperatures based on the biological evidence suggests changes in sea temperatures comparable with the difference existing between northern and southern Australia, probably a little more than

10° C. This curve has been tested by making oxygen isotope paleotemperature determinations on a series of fossils through the Cainozoic taken from southern Victoria within 0.5° of 38.5° S. Series of fossils of the genera (wide sense) *Chlamys*, *Ostrea*, and *Glycymeris* respectively show rises of temperature from lower to mid-Tertiary, then a fall of temperature to the present. The range of temperature is of the order of 10° C. (Dorman and Gill, 1959).

Past Temperatures and Land Faunas

Crocodiles occur plentifully in Australia north of Rockhampton on the east coast and of the Ord River on the west coast. As fossils they occur much further to the south. B. Daily has informed me that four successive faunas are present in the Lake Eyre district of Central Australia, the oldest being the *Perikoala* fauna of Stirton (1957*a*). In all four faunas crocodiles are found. Two fossil crocodilians have been recorded from Victoria on the southern margin of the continent. One came from Oligocene marine beds (Gill, 1961) and the other from subbasaltic fluviatile sediments of Miocene or Pliocene age. Both are remains of large reptiles comparable in size with the living ocean-going *Crocodilus porosus*. The average sea-water temperature for Hobson's Bay (estuarine) at Melbourne in Victoria is 14.4° C. That for Rockhampton is not known to me, but the February (hot season) temperatures in the estuary of the Fitzroy River there are of the order of 28° C. The average for Moreton Bay (estuarine) at Brisbane is 20.9° C. This information suggests that there could be a difference in the mean temperature of the estuarine waters of Melbourne and Rockhampton between 7° and 10° C.

Terrestrial Floras and Temperature

The Australian flora of today is characterized by the genera *Eucalyptus* and *Acacia*. There are hundreds of species of each. However, during the Tertiary, the Australian flora was dominated by broad-leaved types (the so-called *Cinnamomum* flora) and conifers; few of these genera now remain in southern or central Australia, although a number is to be found in the wetter parts of Queensland.

The Araucariaceae (Cookson and Duigan, 1951) may be used to illustrate the temperature changes indicated by the terrestrial flora. *Araucaria* at present extends from the mountain forests of New Guinea, where the average temperature for the coldest month is 64° F. (17.7° C.) to the coastal rain forests of Queensland and northern New South Wales. At the southern limit of the range of *Araucaria* the average temperature for the coldest month is 54.2° F. (18.3° C.). During the Tertiary, however, the pollen species *Araucariacites australis* extended to southern Tasmania (43° S) where the average temperature for the coldest month is 46.6° F. (8.1° C.). This fossil species was originally described from Kerguelen Island in the Southern Ocean, where the average temperature for the coldest month is 35.5° F. (1.6° C.). Thus between the present limit of the genus and the limit of the closely related fossil form, there is a temperature difference of 18.7° F. (10.7° C.).

The Queensland Kauri (*Agathis*) is now limited to northeast Australia, but it is common as a Tertiary fossil, having been identified in many parts of the

mainland, but not in Tasmania. There are suitable ecologies in New South Wales, Victoria, and Tasmania for the Kauri to grow, but temperature is apparently the limiting factor

Summary

Present knowledge indicates an extension southward of the tropic zone during Tertiary time in Australia, with widespread tropical to subtropical forest, and deep kaolinization, bauxitization, and lateritization. Many areas now arid or semiarid then had abundant rainfall. Tropical genera of marine and terrestrial animals extended their range to southern Australia, as also did plant genera such as *Araucariacites* and *Agathis*. The biological (fossil) and physical (oxygen isotope) evidence agree that temperatures rose to a maximum in the mid-Tertiary, then declined to the Quaternary, which was characterized by a series of oscillations. The range of mean temperature for Victoria during the Cainozoic was of the order of 10° C.

References

Bonython, C. W. 1955. Lake Eyre, South Australia—the great flooding of 1949–50. Roy Geogr. Soc. S. A. branch. Adelaide. : 7–9, 27–36, 37–56, 63–68, 69–70.

Butler, B. E. 1956. Parna, an aeolian clay. Australian J. Sci. **18:** 145–151.

Butler, B. E. 1959. Periodic phenomena in landscapes as a basis for soil studies. C.S.-I.R.O. Soil Publ. **14**.

Collins, A. C. 1953. Pleistocene foraminifera from Port Fairy, Western Victoria. Mem. Natl. Museum Melbourne. **18:** 93–105.

Cookson, I. C. 1959. Fossil pollen grains of *Nothofagus* from Australia. Proc. Roy. Soc. Vict. **71:** 25–30.

Cookson, I. C. & S. L. Duigan. 1951. Tertiary Araucariaceae from south eastern Australia with notes on living species. Australian J. Sci. Research. **B4**(4): 415–449.

Dorman, F. H. & E. D. Gill. 1959. Oxygen isotope palaeotemperature determinations of Australian Cainozoic fossils. Science. **130:** 1576.

Gill, E. D. 1955. The Australian "arid period." Australian J. Sci. **17:** 204–206.

Gill, E. D. 1957*a*. The stratigraphical occurrence and paleoecology of some Australian Tertiary marsupials. Mem. Natl. Museum Vict. **21:** 135–203.

Gill, E. D. 1957*b*. The Pliocene-Pleistocene boundary in Australia. Australian J. Sci. **20:** 86–87.

Gill, E. D. 1960. The Holocene history of Australia. Report to I.N.Q.U.A.

Gill, E. D. 1961. The climates of Gondwanaland in Kainozoic times. Chapter XIV. Descriptive Palaeoclimatology. Interscience. New York, N. Y., and London, England.

King, D. 1956. The Quaternary stratigraphic record at Lake Eyre North and the evolution of existing topographic forms. Trans. Roy. Soc. S. Australia. **79:** 93–103.

King, D. 1960. The sand ridge deserts of South Australia and related aeolian landforms of the Quaternary arid cycles. Trans. Roy. Soc. S. Australia. **83:** 99–108.

Mawson, D. 1950. Occurrence of water in Lake Eyre, South Australia. Nature. **166:** 667.

Owen, H. B. 1954. Bauxite in Australia. Bur. Mineral Resources, Geol. & Geophys. Australia Bull. **24**.

Parr, W. J. 1939. Foraminifera of the Pliocene of southeastern Australia. Mineral Geol. J. **1**(4): 65–71.

Stirton, R. A. 1957*a*. A new koala from the Pliocene Palankarinna fauna of South Australia. Record S. Australian Museum. **13:** 71–81.

Stirton, R. A. 1957*b*. Tertiary marsupials from Victoria, Australia. Mem. Natl. Museum Vict. **21:** 121–134.

Whitehouse, F. W. 1940. Studies in the late geological history of Queensland. Univ. Queensland Papers, Geol. **2**(1).

Whitehouse, F. W. 1948. The geology of the channel country of southwestern Queensland. Univ. Queensland Papers, Geol. n.s. **34**.

Whitehouse, F. W. 1954. Artesian water supplies in Queensland. Appendix G. Univ. Queensland Reprints, Geol. n.s. **53**.

Woolnough, W. G. 1927. Chemical criteria of peneplantation (duricrust). J. Roy. Soc. New South Wales. **61:** 17–24.

QUATERNARY CLIMATES OF THE AUSTRALIAN REGION

J. Gentilli

University of Western Australia, Nedlands, W.A., Australia

The purpose of this paper is to arrive at a reconstruction of the climatic conditions of the Australian region immediately before and during the Quaternary Era. Many effects of these climatic conditions are known, but the magnitude of the climatic changes involved and their timing remain unknown. As to their likely cause, a whole literature is available that will not be discussed here. Willett (1949), Charlesworth (1957), and Flint (1947, 1957) give copious bibliographical references.

It is necessary, however, to accept a few basic principles, the first of which is that, on a global scale, climatic changes are symmetrical on both sides of the equator.

Basic Pattern of Climatic Changes

Willett (1953) states that all the available evidence shows that the major glacial-interglacial cycles of change of climate and atmospheric circulation are inphase, rather than in opposition, on the Northern and Southern Hemispheres of the earth. This inphase relation apparently holds also for the secondary cycles of 8 to 10 thousand years' duration superposed on the primary glacial-interglacial cycle.

This confirms the finding by Arrhenius (1950) that there was no significant displacement of the Equatorial Counter Current during Pleistocene time, a fact that appears to be a definite proof that any great climatic changes were "synchronous in the Northern and Southern Hemispheres." The schematic patterns of climatic belts shown by Rubinstein (1956) all show "inphase" (synchronous, simultaneous, symmetrical) variations in the climatic belts of the Northern and Southern Hemispheres.

This rules out the scheme suggested by Taylor (1919, 1927, 1937) and accepted by Keble (1947), according to which all climatic belts would swing north or south respectively with a decrease or an increase of temperature. According to a map drawn by Taylor for temperature conditions equivalent to "present winter and like Pleistocene Equinox," the belt of convectional rains is pushed north of the equator. Evidently a mistake was made in likening the seasonal shift of climatic belts* to the long-term changes in the area and the intensity of these same belts.

It will be shown further that some botanical facts can best be explained with a symmetrical "inphase" change in climatic belts (Burbidge, 1960).

Brooks (1949) discusses the many elements of climate and how each one of these elements could vary with paleogeographical changes. We cannot accept his quantitative assessment of these variations, because it is based on too many assumptions, many of which do not apply or are very hard to apply to Australia. For instance, according to Brooks, an uplift of the land surface

* Modern research clearly shows that even seasonal changes are far more complex than mere "shift of belts" would imply, for example, in monsoonal areas, in the Arctic, and in the occurrence of jet streams.

by 100 feet would decrease temperatures by 0.47° F., while an increase of 1 per cent in the total land area in the latitudes of Australia would increase temperatures by 0.11 to 0.16° F. An increase in vulcanicity would lower temperatures. A more active oceanic circulation, on the other hand, would have a warming effect.

The change in temperature computed by Brooks for each relevant latitude is shown in TABLE 1. This change in temperature refers to the effect of a 1 per cent increase in the land area along the half parallel respectively to the west or to the east of the point in question, for a planetary air circulation such as prevails now. The change in temperature should apply not to the temperature of any one point, but to a "zonal temperature" that, according to Brooks, is a function of latitude and of the proportion of land and sea at that latitude.

All these changes refer to any land mass that remains in its geographical position, that is, disregarding any hypothetical "continental drift."

TABLE 1

Latitude	January		July	
	Land to west	Land to east	Land to west	Land to east
0°	+0.02	0.00	+0.04	−0.02
10° S	+0.07	−0.02	+0.07	−0.05
20° S	+0.13	0.00	+0.04	−0.04
30° S	+0.11	+0.05	−0.02	−0.02
40° S	+0.16	−0.05	0.00	−0.05

Paleoclimatic Effects of Continental Drift

Much has been written about the possibility that continents may have wandered during past geological ages.

Rubinstein (1956) quotes the following supposed latitudes for Perth, W.A., according to the theories of continental drift and polar shift combined: present, 32° S; beginning of Quaternary, 54° S; Miocene, 46° S; Eocene, 40° S; Cretaceous, 70° S.

It will be noticed that between Miocene and the present most substantial changes of climate would result from these changes in latitude alone, even if there were no planetary climatic changes. Fossil evidence does not seem to support the hypothesis, but some theory of continental drift is still invoked by several botanists.

Lam (1934) states that, in so far as New Guinea plants are concerned, there are no relic-endemics known, but there is strong neoendemism, especially at higher altitudes. Asiatic floral elements prevail everywhere, even in the mountains where the climate is cooler; Polynesian ones come next, and Australian-Antarctic ones apparently last, notwithstanding the proximity of Australia, Australian ones being best represented in the mountains and the savannahs. This situation poses a definite problem.

Good (1958) suggests that biogeographical problems of Australia become less perplexing if it is assumed that Australia and New Guinea, instead of being

parts of one and the same primeval land mass, have attained their present proximity in comparatively recent geological times by some relative movement that has gradually diminished the distance between them. Australia may have drifted more or less north-eastward from a position to the south of South Africa. In so doing it underwent a complete change of climates not because of any planetary climatic change but simply by drifting through different climatic belts. According to Good the whole shift may have taken place between the later Cretaceous and the end of the Pliocene.

Mayr (1953) considers that New Guinea and Indonesia may have been on the same continental shelf during the late Mesozoic. Several families of birds may have colonized New Guinea from Asia then. "During the early Tertiary the shelf fractured into a western (Sunda Shelf) and an eastern (Sahul Shelf) component. . . . This rather drastic isolation . . . permitted the development of the very distinct endemic Papuan bird fauna." New Guinea and Australia must have been connected "toward the end of the Tertiary." Needless to say, New Guinea was considerably altered in shape by mountain building.

Various facts stated by Mayr as evidence against the theory of continental drift may be taken as evidence against the drift of New Guinea in relation to southeast Asia, irrespective of any independent drift by Australia. For instance, "the existence of endemic or subendemic tropical Papuan families, such as the birds of paradise, is inexplicable if late Tertiary drift is accepted." Obviously this refers to New Guinea and not necessarily to Australia.

Lam (1934) even accepts the possibility of a Pleistocene "crash" of Australo-Papua into the Moluccan-Melanesian chain islands. Mayr points out that "not a single one of the endemisms among the birds of northern Melanesia . . . more closely related to Moluccan than to New Guinea forms," thus proving that New Guinea must have been part of the Melanesian alignment for a considerable time. The possibility that New Guinea always may have been part of the Moluccan-Melanesian arc, while Australia was quite remote from it, comes within the scope of Good's hypothesis mentioned above. There seems to be some support for this in the botanical evidence, since Burbidge (1960) writes that "it is evident that the relationships with the New Zealand flora lie through New Guinea and New Caledonia, a fact which indicates that the floras of Australia and New Zealand should not be grouped as 'Australasian' unless at least parts of eastern Malaysia are also included."

Good (1958) expects "the Angiosperm fossil record of much of Australia to show in sequence, first a fairly widespread flora predominantly of a temperate rain-forest kind; second, a contraction of this and a spread of an 'Afro-Australian' flora (as exemplified by acacias, Proteaceae, Restionaceae, Ericales, and various Monocotyledons); third, more widespread evidence of aridity accompanied or closely followed by a great increase in such plants as *Casuarina* and *Eucalyptus*, together with the appearance of some development of tropical rain-forest in the northeast and east; and, fourth, one or more short-term extensions of the more mesophytic kinds of vegetation during the pluvial stages of the Pleistocene."

Burbidge (1960) considers that there is "an unsatisfactory basis for accepting the drift hypothesis, especially when affinities with South Africa are so remote."

It is certain that the reconstruction of paleoclimates would be very difficult if continental drift were introduced as a further variable, but at the same time the authoritative acceptance of some such theory and the facts that appear to support it cannot be ignored.

Tertiary Climates

It has become rather fashionable to contrast a "tropical" Tertiary with a "glacial" Pleistocene. The contrast in temperature would be very great, and a general cooling by some 30 or 40° F. would be implied in some of the statements published. Actually the time involved is too short for such a drastic cooling on a world scale; no theory of climatic change can give support to such assumptions, and the amount of cooling likely to have occurred between Tertiary and Pleistocene was probably much less. We may consider some of the evidence put forth so far for a "tropical" Tertiary.

Eocene deposits between Brisbane and Ipswich have yielded remains of such flora as *Cinnamomum*, *Apocynophyllum*, *Ceratopetalum*, *Banksia*, and *Eucalyptus*. Most significant series of similar deposits were found in the Eyrean area, where today are only dry salt lakes and a desert. In New South Wales relics of this "*Cinnamomum* flora" were found as far south as Cooma and Bombala (David and Browne, 1950) but it must be remembered that, during the Eocene, these localities were probably at a much lower altitude than they are today.

Cookson and Duigan (1950) studied pollen deposits from Yallourn, V., including *Banksia*, *Nothofagus*, *Phyllocladus*, *Cinnamomum*, *Tristania*, *Casuarina* and *Callitris*. Cookson in later papers discusses similar deposits from southwestern as well as southeastern Australia; these deposits have been attributed to the Eocene, but a Miocene or even later age has been suggested by other authors.

Balme and Churchill (1959) found abundant *Nothofagus* pollen of at least three species in the Coolgardie depression. Pollens from other plants including *Dacrydium* and *Podocarpidites* varied in quantity, but were mostly rare. In one case fern spores were common, in two cases spores of microscopic algae. Balme and Churchill consider these deposits to be of upper Eocene or lower Oligocene age; "the Eocene sea must have reached at least as far north as Coolgardie during its maximum transgressive phase," that is, about 1500 feet higher than present sea level.

It is possible that a shift of climates may have occurred between Eocene and Oligocene, because fossil remains of similar Oligocene floras seem more frequent farther south. *Cinnamomum* was recorded from Inverell, on the New England tableland (possibly not wholly uplifted then), *Magnolia* from Mount Lofty, S.A., such flora as *Cinnamomum* and *Sterculia* from near Ballarat, V., and *Cinnamomum* and other trees from the Dargo High Plains, V., now at 5,935 feet. In Tasmania *Cinnamomum*, *Apocynophyllum*, *Lomatia*, and *Banksia* were recorded near Launceston and at other places.

Some 30 genera have thus been recorded, among them conifers such as *Phyllocladus*, *Podocarpus*, *Callitris* and *Cupressoxylon*, cycads such as *Mantellia* and *Zamites*, many laurifoliae such as *Cinnamomum*, *Magnolia*, *Tristania*, and *Ficus*.

Much has been made of the "tropical character" of this flora. However, if the presence of *Cinnamomum*, *Tristania*, *Magnolia* and other such genera in these floras can at first sight suggest a tropical environment, the presence of *Nothofagus* can equally strongly suggest a cool temperate environment, and *Hakea* and the Cycads a subtropical one.

Actually *Cinnamomum oliveri*, the Sassafras or Camphorwood of Queensland and northern New South Wales, now grows where no month gets less than 1.5 inches of rain and the mean temperature of the coldest month varies between 54 and 60° F. (Swain, 1928). These conditions are not "tropical" by any means: in fact, Koeppen's limit of tropical climates is a mean of 64.4° F. for the coldest month. Another living species, *C. virens*, has a similar but more restricted habitat (Anderson, 1947). It is true, on the other hand, that *C. laubatii* grows only in the rain forests of northern Queensland (Swain, 1928).

A similar picture would be disclosed by a geographical study of Asian species of *Cinnamomum*, by the geobotany of such flora as *Magnolia* and *Ficus*. Many species are distinctly subtropical, others tropical.

The species of *Nothofagus* were probably similar to those living in New Guinea at the present time. In New Guinea at the present time several species of *Nothofagus*, *Podocarpus*, and *Dacrydium* form the microphyllous montane forests that extend from 6500 to 10,000 feet (Womersley and McAdam, 1957). At these altitudes the mean maximum temperatures vary between 55 and 75° F., and the mean minimum temperatures between 35 and 55° F., the lower temperatures being much more frequent at the upper levels. In Queensland *Podocarpus* grows in areas of plentiful and uniform rainfall, the coldest month's mean temperatures being between 50 and 59° F.; *Nothofagus moorei* is a relic species from a very limited area, above 2500 feet, with a mean temperature of 45 to 50° F. for the coldest month, and a plentiful rainfall (Swain, 1928). Tasmanian species of *Nothofagus* and *Dacrydium* grow in still cooler climates.

It seems reasonable to assume that the Eocene and early Oligocene climate of southern Australia was subtropical and very rainy; there is no basis for any belief that conditions were "tropical." It is the abundance and persistence of the rainfall that made the climate distinctive.

The fact that such plants could grow in the Lake Eyre region, that large trees—still extant as large fossil stumps—could grow in the far southwest of Queensland, points to a very rainy climate. Areas that now receive 5 inches of rain a year must have then received at least 50, with no dry season. If there was even a short dry season, the annual rainfall must have reached some 80 or 100 inches for these laurisilvae to grow as they did. More rain fell in the interior also because the sea extended much farther north: the Nullarbor Plain did not exist at the time and, perhaps a little later, a large part of the Murray Basin was submerged as the Murravian Gulf. Enormous watery expanses occupied the central lowlands, as is shown by the extensive lacustrine deposits of the Eyrean series (David and Browne, 1950).

Pryor (1959) believes that the vegetation of the southwest of Western Australia is distinctively Australian, and that it has been isolated from the east since before the Miocene period. Certainly east-west gene flow was difficult at the time, and remained so to the end of the Tertiary.

Large parts of Australia were subject to prolonged droughts in the Miocene. The extensive lateritic duricrust so widespread in Australia, from the hills of the southwest to the western side of Cape York peninsula, was formed at that time (David and Browne, 1950). However, not all Australia was uniformly affected, and *Cinnamomum*, *Laurus*, and *Sterculia* were still growing near Melbourne.

Evidence of a climate still warmer than at present comes from the few Pliocene deposits. The "newer leads" under some of the Victorian basalt, probably of Pliocene age, contained remains of *Banksia* and *Eucalyptus*, but also fruits resembling those of the rain-forest trees *Owenia*, *Flindersia*, and *Elaeocarpus*.

Thus the "*Cinnamomum* flora" and *Banksia* and *Eucalyptus* were growing in close proximity in the Eocene and in the Pliocene alike. Some authors believe that since most of the fossil remains were deposited in swamps or lakes, we may infer that two floras coexisted, the laurisilvae forming rain forests in the more humid lowlands, and the durisilvae forming more open forests on the surrounding tablelands or ridges. This is quite possible, but it seems most unlikely that such topographic combination should have occurred in nearly every place where these fossils were found. I am inclined to believe that in the Tertiary *Eucalyptus* and *Banksia* were rain forest plants with an exceptional genetic plasticity and physiological adaptability, living in close association with many far less adaptable plants. The structure, form, and carriage of the juvenile *Eucalyptus* leaf is a very significant pointer. It may also be that some of the four or five major sections in which the enormous genus *Eucalyptus* is divided (Pryor, 1959) became genetically isolated during the Miocene climatic changes, while retaining their amazing plasticity.

The Australian Climatic Pattern Today

Australia is the continent with no perennial snow, no mountain climates, and much desert. Australia has no polar continental air and is only subject to occasional incursions of equatorial air; it is the continent most subject to tropical-air influences, to the bright insolation of the tropical latitudes, and to the constant threat of drought. Africa and South America, extending on both sides of the equator, have extensive areas covered by the thick clouds of the intertropical convergence zone. In Australia the intertropical convergence does not go very far south. The presence of a body of water along its northern shores is of no climatic advantage to Australia; in fact it may be interpreted as weakening the ITC and even impeding its southward progress during the hot season.

At present northern Australia has much more marked continental characteristics than the much bigger South America. These characteristics are most pronounced in the summer maxima and in the winter minima. All this makes one wonder whether the "Australian monsoon" has been grossly overrated, especially since monsoonless continents have a better rainfall than Australia which in the same latitudes is noticeably hotter.

The dominance of summer winds with a westerly component is not absolute over northern Australia except between 125° and 135° east. It is westerly and southwesterly winds that predominate, that is, not air streams from the other side of the equator, but deflected southerly air streams, some of which have

traveled over the interior of the continent before being deflected around and into the stationary beat-low that forms over northwestern Australia at this time of the year. Northwesterly winds, of true monsoonal origin, are important only in the far north and even in Darwin in summer are not as frequent as southerly winds. At Darwin, in January, westerly component winds occur 72 per cent of the time at 500 m., 66 per cent at 1000 m., 49 at 2000, 29 at 3500, and never at 5500, where easterlies occur 85 per cent of the time.

Thus, during November and December, westerly winds are frequent at low levels, but they are very shallow. In January and February, westerly winds become much more prominent at low levels and up to 3000 m., but the northerly component is relatively uncommon. The true monsoonal stream usually ends at about 9° S.

Rain is recorded in less than 10 per cent of the December–February observations all along the northern Australian coasts with the exception of the Kimberley coast where orographic factors trigger off about 80 to 120 thunderstorms a year, and where observations of rain increase to 10 to 15 per cent. Obviously the frontal zone, not a very active one because of the almost parallel streams that cause it, is usually outside the Australian continent.

In winter the trade winds (tropical easterlies) dominate the weather and conditions are very dry, but not as completely dry as in Africa at the same latitude.

Conditions are very dry from May to September, and very wet from December to March. If one takes a value of R/E = 0.5 as the limit of effective rainfall one finds that effective rainfall is available from the end of November to the end of April, little over 5 months. The ratio for the yearly totals is about 0.8.

Little further south the rainfall decreases very rapidly, and the evaporation increases, so that at Mataranka the number of months with ineffective rainfall rises to nearly 9, and that of months with effective rainfall drops to little over 3.

Most of the rain actually comes in heavy downpours, and for this reason the rainfall incidence per wet day is very high. Infiltration of rain water into the soil is much less thorough than farther south, and run-off is greater. For this reason an R/E ratio of about 0.5 is considered as effective a ratio as 0.3 in southern Australia.

Northern Australia, characterized by a short rainy high-sun season followed by a long dry season, may be divided into climatic regions according to the number of months of effective rainfall. A 3-month effective season may be the shortest season for tree growth under normal conditions, and this is about the limit of the Australian woodland savanna. The region farther south, with a shorter effective season, belongs to the dry interior.

Only a short tract of the coast is subject to the trade winds the whole year round, or nearly so, from Cooktown to Bundaberg. Farther south the traveling anticyclones bring variable winds, among which southeasterlies still abound, but are by no means dominant, except in summer as far south as Brisbane. Between December and February the heating of the interior causes some deflection of the winds, which assume a more westerly component, but may still be called modified trade winds, because the modification does not warrant the name of "monsoon." At Cooktown for instance, easterly-component winds occur in

45 per cent of the mornings and 73 per cent of the afternoons in January, when westerly-component winds occur in 41 per cent of the mornings and 20 per cent of the afternoons. Since land breezes come approximately from the west and sea breezes from the east, it may be said that in January a trade-wind element is noticeable in at least 45 per cent of the observations, and a monsoon element in at least 20 per cent. Even when one allows for the variation in sea-breeze direction and for the short period of observations (only 5 years in most localities) one notices the gradual shift from summer westerlies in the north to winter westerlies in the south, and the fact that summer westerlies become uncommon south of Cairns. The name of "trade-wind coast" may therefore be properly given only to the coastal belt from Cooktown to Bundaberg, where tropical easterlies dominate the weather. In a typical trade-wind coast, undisturbed by continental-monsoonal influences, the rainfall varies very little throughout the year. Along the Queensland coast there is a definite trend, with the rainfall becoming more evenly distributed toward the south, but still remaining outstandingly uneven all along the trade-wind coast, which obviously is not typical.

The general trend of the coast is parallel to the prevailing wind direction. When the trade winds become definite easterlies (when they are deflected toward a continental low-pressure area in summer) they cut the coast and bring plentiful rains, but this happens only part of the year. Where the coast has a meridional trend, through Cairns to Ingham, the angle of wind incidence is wider and the rainfall more even. South of Ingham the coast is again parallel to the wind and the seasonal proportion of the rains follows the previous trend. At Rockhampton or Gladstone the rainfall may be called uniform, by conventional standards. Thus, just where the trade-wind coast ends, the more even rainfall begins. In this respect too, Australia is probably unique.

Mareeba, the tobacco-growing center on the Atherton Tableland, has an effective season of 5 months, and so has Townsville, which is one of the world's driest localities on the trade-wind coast. Only a few miles farther inland the dry climates begin. On the other hand, Innisfail has no dry season, and Rockhampton, at the edge of this climatic belt, already has a 10-month effective season.

Most of Australia is in the belt of tropical divergence under the control of subtropical traveling anticyclones. These anticyclones consist of enormous bodies of falling air (from about 30,000) feet that rotate anticlockwise and spread out near the ground. Because of its relative dryness the rate of warming-up is rapid, and high temperatures are common, especially when summar insolation can heat the underlying land mass. Outgoing radiation is also effective, and night and winter temperatures are relatively low. Over most of Australia, where anticyclones are prevalent, the mean daily range of temperature varies between 25 and 30° F., whereas it is only 9 to 14° F. in the ITC belt and in the belt of westerlies. It is only on the north side of the anticyclones that the winds gradually steady onto a northwestward course to become the trade winds. Within each anticyclone the successive directions of surface winds are regular and well known; between anticyclones they are conflicting, but nearly so in summer, with a quasi-stationary front and light clouds, more irregularly in

winter, with multiple cold fronts and stormy weather. On the south side of the anticyclones the westerlies appear, irregular and often violent.

The dry regions have only one common characteristic: inadequate precipitation. In the north they get occasional rain from thunderstorms, in the east from the trade winds, in the west and south from westerlies and cyclones, here and there, but once in a few years perhaps, from a tropical cyclone, but the rain is irregular and unreliable. It varies considerably. At the northern margin of this anticyclonic belt there are occasional and erratic summer rains, at the southern margin light winter rains, over an immense continental area no definite rains save from occasional far-reaching tropical hurricanes or from erratic local thunderstorms.

Over the windward coastal belt however, trade winds and anticyclonic winds bring enough vapor to cause rains, especially where they meet with an orographic lift. The tilted-block build of the eastern highlands, with the steep side facing east, favors a very heavy rainfall over the eastern seaboard, and a much lighter but far more widespread rainfall toward the interior. The winter rainfall generally is somewhat heavier and more irregular because of the warm East Australian Current that flows past the shore.

From Rockhampton or Bundaberg southward, the coast is characterized by an increasing frequency of winter westerlies, although the main climatic factor is the succession of traveling anticyclones.

As far south as Moruya Heads the westerly component is dominant during the winter months, but farther south it becomes dominant throughout the year. The subtropical east coast extends therefore from Rockhampton or Bundaberg to Moruya Heads or a little farther. North of this limit the daily amplitude of pressure is usually greater than 2 mb., south of it is well below 2. Along the subtropical east coast the mean annual temperature is generally above 29° F. and the mean daily maximum above 36° F. Farther south the mean values are lower.

For a number of months the actual rainfall is only just above the effective quantity, but by so little that in any dry year it becomes ineffective. Tamworth is outside the forest region, and we may assume that the limit of the subtropical east-coast region lies somewhat east of Tamworth, where the sum of the 12 monthly R/E ratios reaches about 9. Farther inland the climate is subhumid, and at Coonamble conditions are semiarid. Cobar lies definitely within the arid region.

As in the corresponding regions of other continents, the subtropical west coast is characterized by an alternation of variable anticyclonic winds in summer and westerlies in winter. The anticyclonic winds are relatively dry because they are katabatically heated, and even when they finally blow over an expanse of water and absorb more moisture, they do not precipitate it on very hot land, where their relative humidity suddenly decreases.

The peripheral anticyclonic stream in its southeastern quadrant is maritime at Fremantle (west coast) and continental at Newcastle (east coast). In its northwestern quadrant it is continental at Fremantle and maritime at Newcastle. Hence, other things being equal, the summer anticyclones bring northeasterly continental air over Fremantle and northeasterly maritime air

over Newcastle. Northeasterly winds in the southern hemisphere come from lower to higher latitudes and therefore increase their relative humidity, but the high temperatures of the dry interior are such that saturation point is very seldom reached. On the other hand saturation point is more easily reached on the eastern coast, after the winds have traveled over broad expanses of ocean. Southwesterly winds travel equatorward, and their relative humidity becomes lower; when they hit the hot Australian coast they fail to yield any rain, although there may be a stationary front and some clouds, especially before and after summer. Thus Fremantle hardly ever gets any rain from peripheral anticyclonic winds in summer, while Newcastle gets some good rains.

In winter the whole system has moved north, and any maritime air, on reaching the colder continent, may release rain; but the strong northwesterlies of the middle latitudes are much more effective on the west coast than northeasterlies can ever be on the east coast, hence the difference in the winter rainfall of the two.

Like Eurasia, Australia affords a great longitudinal development to her "mediterranean" climate, but whereas in Syria, Lebanon and Israel the mediterranean climate merges with the dry climates, in Victoria it merges with the east-coast climate, a situation which is unique on such a scale (there is a small occurrence of this transition in South Africa).

This great longitudinal development causes a gradual modification of the proportion of winter rains, from the very high proportion found at Bookara and Geraldton to the practically uniform rainfall in western Victoria.

Situated on the poleward side of the subtropical regions are the cool temperate regions, where the hottest summer month is below 22° C. and the rainfall is uniform. The prevailing winds are westerlies, but it should be noted that altitude makes the region extend north into subtropical latitudes, where more easterlies prevail. The westerlies travel with the succession of cold fronts so common in these latitudes in the Southern Hemisphere, and the result is a rainy climate.

Because of the latitudes where this region is found, there is a gradual change from a summer maximum in the north to a marked winter maximum in the center and more uniform conditions in the south. In Tasmania there is a striking difference in quantity but not in regime between the west coast and the east coast. Throughout the region, however, there is no dry season. The effective season is very long in the south, and would be continuous were it not for the ineffective temperature of some part of the year.

Only very small areas in the Australian Alps and on the Tasmanian Plateau have low temperature effectiveness, and hence a cold climate.

Climatic Peculiarities of the Pleistocene

Following upon the decision by the 1950 International Geological Congress the beginning of the Pleistocene was set to include the Calabrian (marine) and Villafranchian (terrestrial) formations. Gill (1957) proposed to set the beginning of the Pleistocene in Australia at the base of the Werrikoo formation (Victoria) "in which fossil assemblages indicate climatic conditions similar to the present." Earlier deposits show evidence of a warmer climate.

Flint (1947) pointed out that since, during the glacial ages, the regional snow ne was lowered roughly parallel to itself, and not more on one side than on the emaining sides, it must be inferred that lower temperature and not increased recipitation started each glacial age.

It is generally accepted that climatic fluctuations were greater in the higher titudes, but it is obvious that in glaciated areas much heat would be lost to pace because of the greatly increased albedo of snow-covered ground, and that uch of the absorbed heat would be spent in melting ice, without a rise in emperature (Charlesworth, 1957).

A general fall of temperature is assumed at all seasons, but especially in ummer. Equatorial temperatures were about 7° F. lower than at present. he depression of the snow line in tropical regions was about 1600 feet, cor-esponding to a drop in temperature of only 4.5° F.

In mid-latitude regions the snow line descended by over 2200 feet and there ay have been a greater fall in temperature, the more widely accepted value eing between 10 and 12° F.

The glacial climatic pattern "is characterized by a considerable expansion of ne polar cap zone of glaciation and by a corresponding equatorward shift and titudinal compression of the present zonal pattern. The middle-latitude elt of maximum cyclonic rainfall probably extends its influence 15° farther quatorward into the present subtropical high-pressure belt of relatively dry nd settled weather. The zonal westerlies . . . are considerably expanded and so intensified . . . with increased storminess." (Willett, 1953).

There is "a marked acceleration of the condensation cycle . . . manifested . . . so in the extremely pluvial conditions that prevail in the lower middle lati-udes (which today experience semiarid subtropical conditions), and in the eatly increased rainfall in the belts of equatorial convergence between in-nsified tropical easterlies."

This generally strengthened circulation gives rise to "increased subsidence d divergence in the narrowed and intensified subtropical high-pressure elt . . . in which the evaporation phase of the accelerated condensation cycle ust be maintained at high efficiency."

The interglacial climatic pattern is more or less the opposite, with the pole-ard shift of the westerly cyclonic belt, and "a great increase of warmth and ecipitation in the polar latitudes." The condensation cycle is slower, and he entire zonal pattern of climate is diffused and weakened." It seems also ely, as Kraus (1960) agrees, that glacial periods were "characterized by a more tive conversion of thermal into mechanical energy." A strengthening of the opical easterlies would bring a heavier rainfall to the equatorial belt. The oper westerlies would also probably be stronger and closer to the equator, rticularly in summer. Meridional troughs and depressions would be frequent. owever, as Kraus points out, "not enough is known about the relation be-een the circulation pattern and the distribution of heating and cooling . . . ; e probable coincidence of pluvials near the equator with glacial conditions in gher latitudes . . . cannot be accepted as proven until it has been confirmed additional geological evidence."

Much has been written on the likely causes of glacial periods. Some of the

hypotheses may lead to significant differences in the climatic conditions likely to have affected Australia.

According to Rubinstein (1956), an increase in solar activity (as postulated by some authors) would cause:

(1) An increase in the latitudinal width of the equatorial-tropical zone, which would push the subtropical high-pressure belts farther polewards.

(2) The limits of the polar zones would be pushed into lower latitudes, and the planetary circulation would be strengthened ("high-index" circulation) but polar winter temperatures would be warmer.

(3) The temperate belts would become narrower and their climate less continental; the latitudinal gradient would be greatly reduced.

(4) With the strengthening of the planetary circulation the temperature of the equatorial and tropical zones would drop slightly because of the more active heat transport toward the higher latitudes and because of the greater cooling of huge convectively rising air masses.

An entirely different theory, propounded and abandoned and revived again, attributes many a climatic change to variations in the composition of the atmosphere, especially in its content of carbon dioxide (Plass, 1956).

Kraus (1960), accepting this theory, stresses the importance of water vapor, carbon dioxide, and ozone, factors that determine the upper limit of convection (tropopause) for any given temperature, because they control the rate of infrared radiation cooling of the earth's surface. A tabulation of these variables may be useful: Low latitudes (O_3 may often be important): (1) low CO_2 and/or low O_3 mixing ratio (m.r.) causes high tropopause, stronger easterlies, more intense convectional cooling, towering CuNi, heavier rainfall; (2) moderate CO_2 and/or O_3 m.r. cause moderate easterlies, lower tropopause, higher temperature, decreased rainfall; and (3) high CO_2 and/or O_3 m.r. causes very weak easterlies, extremely low tropopause, frequent storms, plentiful fogs but scarce rainfall.

Middle latitudes (O_3 may occasionally be important): (1) low CO_2 (and/or low O_3) m.r. cause high tropopause, higher zonal index (stronger westerlies, jet stream) and increased cyclogenesis, heavier precipitation, prevalent oceanic conditions; (2) moderate CO_2 (and/or O_3) m.r. cause low zonal index (blocking, intermittent westerlies), alternation of troughs with anticyclones, greater continentality.

High latitudes (O_3 unimportant): (1) low CO_2 m.r. favors radiant heat loss and glacial surface temperatures and, in continental areas, stationary anticyclones and scarcity of precipitation; (2) moderate CO_3 m.r. reduces radiant heat loss and favors moderate surface temperatures.

Thus during a glacipluvial phase the following climatic changes would affect Australia, irrespective of the initial cause of the change in global climates: (1) the rainy equatorial belt would spread southward and, probably, a south intertropical convergence would affect northern Australia; (2) southern Australia, from the southwest to Victoria, would be under the constant influence of the westerlies; (3) its climate would be much wetter and distinctly oceanic; (4) the anticyclonic belt would be very narrow and interrupted, and its climate would be semiarid on the average, but with extremely arid "spells"; (5) the "mediterranean" climatic wedge would be almost nonexistent.

It follows therefore, because of points 1 and 2 above, that a pluvial phase vould affect both northern and southern Australia at the same time.

There were also secondary climatic cycles of 8 to 10 thousand years' duration hat appeared in superposition on all four of the Pleistocene glacial maximums lso inphase in both hemispheres (Willett, 1953).

Australian Pleistocene Climates

These climates are often surmised from the nature of fossil remnants found t various places. It should be stressed that the evidence of these fossils may e misinterpreted because of changes in the habits and habitats of species and, specially, because Australia was large enough to have different climates at the ame time.

A very important factor that contributes an element of doubt to every at-empt to reconstruct past climate from the biological (paleontological) record s the variability of somatic and other characteristics. For instance, contem-orary observations show that lake and pond mussel populations "are usually nore swollen, with the shell outline more rounded and relatively higher, or vith the posterior end expanded and with lighter shells than their river-dwelling ounterparts. Populations in fast-flowing rivers usually show more elongate hells with straight or sinuate ventral margins, less swelling and heavier teeth.... oose mud appears to be correlated with lightness of shells and posterior longation, whereas coarse gravel is associated with heavier shells shortened osteriorly.... Such ecologically adapted populations (ecotypes) are com-non... and such ecological adaptations may eventually become incorporated nto the genotype of the species" (McMichael and Hiscock, 1958).

How rapidly a change in thermal conditions can act as a selective agent is hown by the experiments of Boden (1958) with *Eucalyptus fastigata*. Seeds vere collected from the same mountain slopes at 2350, 3500, and 3850 feet espectively, and planted under uniform conditions in the open at Canberra. Exposure to frost killed more than one half the low-altitude seedlings, while nost of the high-altitude seedlings were not even damaged. This is probably vhy Pleistocene climates are so difficult to interpret biogeographically.

Throughout these notes the European terminology (Günz, Mindel, Riss, Vürm) is used.

The preglacial stage. The preglacial stage of the Pleistocene is believed to ave begun 825,000 to 660,000 years ago by Zeuner, at well over 1,000,000 ears ago by Flint. Crocker and Cotton (1946) and Tindale (1947) linked it vith the Naracoorte marine terrace of South Australia (220 to 250 feet above ea level). Jennings (1959) found some remnants of steps at 225 to 238 feet n King Island, Davies (1960) noticed a series of remnants at 240 to 270 feet n the Tasmanian coast.

A large part of the Murravian Gulf still persisted at this time, and much of he Nullarbor Plain was still under water, as is shown by fossils at Ooldea, S.A., ow at 380 feet above sea level.

The Mallee Ridge rose slowly and dammed back the Murray, forming Lake Tawait, which probably covered about 8,000 square miles (David and Browne, 950). The Kosciuskan uplift may have continued for some time.

Climatic conditions were probably warmer and more humid than at pres-

ent. A fossil crocodile from Port Augusta, S.A., may date from this time (David and Browne, 1950).

The Günz glacial. The earliest known Pleistocene glacial (the Günz period of European chronology) dates back between 590,000 years (Zeuner) and more than 1,000,000 years (Flint). No trace of glaciation from this period subsists in the Australian region so far as we know. In the absence of data we must assume that, as in Northern-Hemisphere evidence, the intensity of glaciation in this stage was about the same as in the Würm glacial. The snow line is now at about 8000 feet above sea level in Southern Australia (Süssmilch, 1941), and at about 15,400 feet above New Guinea (Reiner, 1960). We may suspect a depression of the snow line of over 2000 feet, corresponding to a cooling of about 6° F. in the mean annual temperature. The winter snows at 6000 feet on Mount Kosciusko have been known in recent years to accumulate occasionally to a depth of 25 feet, with an equivalent water content of more than 100 inches. However, spring and summer temperatures are so high and radiation is so strong that the snow cannot last through the summer. An increase of precipitation sufficient to do so with the present temperatures would have left some geological evidence and physically, would be hardly possible, but prolonged general cloudiness would have made a difference. On the other hand, a drop of about 5 to 7° F. in the mean summer-air temperatures would allow extensive snowfields to last through some summers, and it might permit the re-establishment of glaciers even if the present radiational heat gain remained the same (Kraus, 1960).

We may tentatively suspect a slightly colder winter and a definitely cooler summer than at present, some glaciation on the Tasmanian mountains, a long belt of humid climate in the south, interrupted by a much more pronounced Great Australian Bight, with westerly winds and frontal rains almost throughout the year. The "mediterranean" climate would be restricted to a very narrow triangle on the western coast, perhaps between Geraldton and Sharks Bay. A belt of semiarid country would occur west of the Musgrave and Macdonnell Ranges, which would be cool and rainy, with rains predominantly from the western quarter.

In the east, westerlies and frontal rains would control the weather as far as Brisbane in the winter, as far as Sydney in the summer. Variable, dry anticyclonic winds would occur between Geraldton and Brisbane in the summer and, occasionally, nearly along the tropic in the winter. Strong easterlies would blow over northern Australia in the periods of anticyclonic weather, but broad wedges of equatorial air would interrupt them and bring torrential rain.

The Günz/Mindel interglacial. During the Günz/Mindel interglacial the sea rose (Milazzian transgression of 184 ft.) and formed a series of terraces known as the Cave Range, S.A., at 180 to 190 feet (Crocker and Cotton, 1946; Tindale, 1947). The time may have varied between 500,000 (Zuener, 1945) and 900,000 (Flint, 1947) years ago.

Climatic conditions were probably warmer than at present (David and Browne, 1950).

The Mindel glacial. Still early in the Pleistocene came the Mindel glacial of Europe, considered by some authors (for example, Tindale, 1947) to correspond

to the "Malanna glacial" of Tasmania. The time would have been 450,000 (Pfannenstiel) or 476,000 (Zeuner) to 700,000 (Flint) years ago.

A marine (Roman or Pretyrrhenian) regression of over 250 feet may have occurred (David and Browne, 1950) but this is by no means a widely held belief.* Similarly, Lewis (1934*b*) does not think that a global correlation can be established and prefers to leave his three Tasmanian glacials uncorrelated. Flint (1947, 1957) correlates them with glacials later than Mindel. On the other hand, David and Browne (1950), on the assumption that "Malanna" and Mindel were correlated, considered this period locally frigid enough to bring the snow line to nearly sea level in Tasmania and to about 4800 feet on Mount Kosciusko, a difference that seems too great.

Jennings and Banks (1958) consider that nothing is left of any early Tasmanian glaciation.

The position and characteristics of climatic belts would not have differed greatly from those outlined for the Günz glacial.

The Mindel/Riss interglacial. Tindale (1947) links the Mindel/Riss interglacial, which lasted about 185,000 years, and its Tyrrhenian marine transgression with the East Avenue Range (105 to 110 feet) of South Australia. Crocker and Cotton (1946) had also included the Baker Range (140 to 145 feet). David and Browne (1950) mention signs of submergence to 100 to 150 feet. Davies (1960) found poorly developed but fairly clear terraces at 120 to 150 feet in Tasmania, Jennings (1958) steps at 127 to 144 ft. on King Island.

David and Browne (1950) consider this period probably very much warmer than the present. Remains of the crocodile *Pallimnarchus pollens* found in Queensland near the Warburton River and near Port Augusta, S.A., indicate a heavy rainfall, adequate to create rivers, and a hot climate.

The Riss glacials. The Riss I glacial of Europe, dating back 235,000 (Zeuner) to 325,000 (Flint) years ago, was by Tindale (1947) linked with the "Yolande I" stage of Tasmanian glaciation, but recent work (Jennings and Banks, 1958) shows that Tasmanian glacialism needs revision.

An emergence of 305 feet because of the lowered sea level was postulated by Antevs (1928), but Daly (1934) considered it more likely to have reached about 294 feet. Davies (1960) mentions traces of a lower sea level at −210 feet on the Tasmanian coast. He also found channels cut into bedrock by rivers near their mouths to at least 180 feet below present sea level.

Since this phase may have lasted a very long time (David and Browne, 1950) it may be expected that considerable downward erosion by rivers should have taken place.

David and Browne (1950) consider this period generally cold and wet, locally frigid. Glaciation certainly occurred in Tasmania and on Kosciusko.

There undoubtedly was extensive glaciation in Tasmania at this time, but Jennings and Ahmad (1957) and Jennings and Banks (1958) have failed to find any relic from it. As mentioned above, the three phases propounded by Lewis, 1934*a*, (Malanna, Yolande, Margaret) need revision and it may prove necessary to abandon them.

Browne (1945, 1952) recognizes three "episodes" in the Kosciuskan glacia-

* A regression of up to 400 ft. has been mentioned.

tion. The first was a small icecap (questioned by Galloway *in litt.*) that descended at least to 4800 feet above sea level; the next was characterized by valley glaciers; the last one by small cirque- and valley-head glaciers. "Whether these were all phases of one cycle of glaciation or were independent and distinct and separated by long time intervals is a debatable question" (Browne, 1952). The icecap stage may date from the Riss glacial.

Controversial traces of glaciation may occur on Mount Bogong, V. (Carr and Costin, 1955; Beavis, 1959). If there was only one Pleistocene glaciation in New Guinea, it is also likely to have been in this period.

Mount Carstensz (16,503 feet) lies in western New Guinea, in a region where the monsoonal reversal of wind direction is very pronounced. Also, southeasterly winds reach the area after having drifted over southern New Guinea and, at times, northern Australia. Their moisture load is much less than farther east, and the May to October rainfall is accordingly moderate. The present snow line on Mount Carstensz is set by Dozy (1937) at 15,345 feet, but the altitude of cirque floors is 13,400 feet and that of cirque lakes (as interpreted by Reiner, 1960) is 12,500 feet, which may also be taken as the snow line. Pleistocene depression of the snow line on Mount Carstensz may thus have varied between 2000 and 3100 feet.

On Mount Wilhelmina (15,580 feet) a lake, striated boulders, and rock ledges have been observed at 13,300 feet of altitude (Lorentz, 1911).

On Mount Wilhelm (14,900 feet), in eastern New Guinea, snow falls only once or twice a year, chiefly in August. The whole summit area may then receive a snow cover, which lasts from a few hours to two days. Smaller patches of snow may persist longer in sheltered positions close to the summit (Reiner, 1960). There is no perennial snow. Reiner (1960) places the present theoretical snow line at 15,400 feet.

Cirque lakes are found between 11,500 and 12,000 feet, and the Pleistocene snow line may therefore, according to Reiner, have been depressed by 3400 to 4000 feet in comparison with the present.

If snow is so scarce and infrequent on Mount Wilhelm, at 14,900 feet now, it is hardly likely that under the same conditions it should be perennial only 500 feet higher up. Have the winds changed? Of some 53 glacial cirques shown on Reiner's (1960) map, only 10 are north of the main divide; 36 face south, southeast or southwest, and 12 in all face north or northwest. Although the steepness of the slopes may have been a selective factor in Pleistocene snow cover, the prevalent direction of the moisture-bearing winds was probably much more important. In fact, even north of the divide there are more cirques on the south-facing than on the north-facing slopes of the valleys. The 2 north-facing cirques north of the divide are so close to the summit ridge that some of the snow that formed them could have drifted from the ridge itself; also, they both occur where the main crestline runs north-south, thus intercepting any southeasterly winds.

Is the difference in the Pleistocene depression of the snow line between Mount Carstensz and Mount Wilhelm really significant? At the present time the snow line lies at about the same altitude in both areas (Reiner, 1960) or, if there is any difference, it is likely to be higher above Mount Wilhelm. In

the Pleistocene it was higher (by as much as 500 to 1900 feet) on Mount Carstensz than on Mount Wilhelm. This probably means that, in the Pleistocene, Mount Carstensz was relatively drier and that this, in turn, was probably due to greater continentality, caused by the emergence of the vast continental shelf. Mount Wilhelm, much farther east, was then, as now, exposed to the easterlies from the Coral Sea.

According to Reiner (1960) occasional frosts occur at the summit of Mount Wilhelm. In the Pleistocene it was likely that occasional frosts occurred at about 11,000 feet, and the daily mean temperature at that altitude (computed according to the usual lapse rate of 3° F. for every 1000 feet) would have been about 47° F. Daily maximum temperatures would have varied between 52° and 62° F.; nightly minimum temperatures between 32° and 42° F.

Not far above 11,000 feet conditions would have been ideal for the peculiar kind of nivation known as *nieves penitentes*, pinnacles (hence the reference to the white-hooded *penitentes* of some Spanish processions) due to the daily thermal oscillation about freezing point (Troll, 1942).

Near the summit, the daily mean temperature at the time of maximum glaciation would have still been above freezing, probably around 35° F. Almost every night would have seen temperatures well below freezing, and glacial phenomena must have resembled those of the northern Andes under present conditions.

Discussing the biogeography of Queensland Echinoderms, Endean (1957) agrees with Fairbridge (1953) that an emergence of less than 60 feet would close Torres Strait. A major barrier to marine biogeographical migration would thus arise. It is very likely that closely related species would "have arisen from geographical isolates which initially diverged genetically during the Pleistocene lowering of sea-level." Endean (1957) lists 12 pairs, 4 triads, and 2 larger series of species of marine Echinoderms which are now sympatric, because their ranges have probably spread and overlapped after the disappearance of the former Torres Isthmus barrier. On the other hand, 7 pairs of species have remained separate (allopatric).

The present mapped distribution of the freshwater mussels *Velesunio* and *Westralunio*, closely related genera, shows *Velesunio ambiguus* in the Murray-Darling basin and in the rivers of eastern Queensland, *V. wilsonii* in the Lake Eyre system and in several rivers of eastern Queensland, *V. angasi* in the rivers of Carpentaria, Arnhem Land and the Kimberleys, and *V. sentaniensis* in northern New Guinea. On the other hand *Westralunio carteri* is found in the southwest, and *W. flyensis* in southern New Guinea (McMichael and Hiscock, 1958).

We interpret these patterns as follows:

(1) In pre-Pleistocene times, *Velesunio* inhabited the Eyre and Murray-Darling drainage systems; *Alathyria*, the eastward-draining rivers; *Westralunio*, the rivers of the western shore; and *Microdontia*, part of New Guinea.

(2) *Velesunio* seems to have spread in several waves. From the Eyre system an earlier population of *V. wilsonii* colonized the Carpentarian rivers, and, after a period of segregation, became specifically different (*V. angasi*). *V. sentaniensis* may have become separated by orographic uplift. From the

Murray system *V. ambiguus* colonized northern Tasmania, evolving into the *V. legrandi* of today.

The regression of sea level must have provided causeways to Kangaroo Island, to the islands of Bass Strait and Tasmania and to New Guinea.

On Kangaroo Island, S.A., fossil remains of giant marsupials (*Diprotodon, Nototherium, Sthenurus*) were found in what probably were early Pleistocene deposits (David and Browne, 1950).

Jennings (1959) recalls various finds of marsupial remains from King Island: *Nototherium tasmanicum* on an islet in a small lagoon; *Diprotodon optatum* in clay from the same lagoon; *N. mitchelli, N. victoriae, Macropus anak,* and *Zaglossus harrisoni* from a drained swamp; *M. anak, Sthenurus atlas,* and remains of species extinct on the island but still living on the mainland, from blowouts. The large size of these mammals makes it certain that they must have reached King Island when a land connection with the mainland was available. It is also likely that the connection may have persisted for a long time, since these remains do not differ appreciably from the corresponding ones from the mainland.

Fleming (1942) analyzes the evolutionary sequence and the geographical distribution of the prions, or whalebirds. He points out that "morphological divergence has been engendered by physiological isolation due to differences in sexual periodicity, which are, in turn, in some way linked with the different environmental periodicities of the zones of surface water over which the birds live." In the early Pliocene a general climatic cooling would have caused a first split in the prion population, with consequent speciation. In the late Pliocene the East Australian (Notonectian) Current began to appear, pushing warm waters farther south, and causing the separation of the two species of *Pseudoprion.* In the Pleistocene a cold phase caused a northward shift of water convergences; thus the subgenera *Pachyptila* and *Heteroprion* became separated from the main stock and differentiated. In an interglacial phase, specific differences began to appear with each subgenus.

Thus only one glacial and one interglacial phases are invoked.

Fleming (1958) shows that the New Zealand scallop *Pecten benedictus*, found in the lower Pleistocene in a distinct fossil form *P.b. marwicki*, differs from the upper-Pleistocene *P.b. tepungai*. However, immediately below the upper level another fossil form occurs, which is clearly intermediate in height/length ratio, in frequency of lamellae on flanks, in right-valve inflation, between *marwicki* and *tepungai*. The concept of cline, usually referred to spatial sequences, must also be referred to chronological sequences.

Pecten benedictus is now found in southeastern Australia, Hawaii, the Yellow Sea, and the Red Sea, all warm areas. Is it possible that the somatic evolution of the New Zealand Pleistocene forms was due to changes in the environment? The emergence of large land areas north of Australia during the glacial periods might have deflected more warm water southward to New Zealand, thus causing an apparent contradiction: the flourishing of a warm-water species in glacial times.

A thorough review of the phytogeographical constitution of the Australian continent is made by Burbidge (1960). We shall quote the main points relevant to a paleoclimatic reconstruction.

Genera with discontinuous areale occur in the north; they have strong Malaysian affinities. It may be inferred that they migrated from Malaysia during a marine regression, attained a wide distribution during a humid climatic phase, and had their areale subsequently fragmented by the spread of a drier climate. This relatively recent fragmentation has led to speciation.

During the Riss glacial, conditions in Australia were rather cold and humid. The westerlies were then much farther north than at present, giving rise to a cool rain-forest habitat extending across Australia from Perth to Sydney and along the Eastern Highlands to north of Toowoomba. Forest habitats extended over a much larger area than at present, and had outliers in the mountains of Central Australia.

I cannot agree with Keble (1947) who, accepting the general suggestion by Taylor (1919, 1927), believed that the climatic belts moved northwards in glacial periods, moving with them the forest, savannah, arid, and steppe belts. A variation in temperature of 12° F. was involved, equivalent to a movement, north or south, of 800 miles. In a previous paper (Gentilli, 1949) I considered a similar migration of climates, but with the arid and steppe belts slightly contracted.

However, bioclimatic evidence seems to show that the increases in rainfall of any glacial period may have occurred both north and south of the continent. For instance, Burbidge (1960) finds some indication of the fact that both northern and southern Australia were pluvial at the same time in the tropical affinities of the northern Eremean species. She finds no evidence of any past invasion of northern Australia by southern Eremean plants, as would have taken place if all the climatic belts had moved northward unchanged. The museum scientist Keast (1959) gives a lively account of the geography of relict plants and animals in the Macdonnell Ranges, Central Australia. An interesting aspect of the well-known distribution of *Macrozamia* and *Livistona* is that while *Macrozamia* has southern relatives, most of the *Livistona* species are from northern Australia, the only exception being *L. australis*.

We may infer that in some earlier period, probably Pleistocene because of the speciation involved, humid climates suited to *Macrozamia* extended right across southern Australia, at least as far north as the Macdonnell Ranges. At some (other?) period humid climates suited to *Livistona* extended right across northern Australia, at least as far south as the Macdonnell Ranges.

The same museum scientist (Keast, 1959) mentions the freshwater fish *Fluvialosa richardsoni*, as confined to permanent rock holes in the mountains of central Australia. The closest relatives of this species are in the Murray-Darling system. A cap limpet (*Stimulator consetti*), collected in the Macdonnell Ranges, was formerly known only from the Harding Range in northwestern Australia (D. F. McMichael, in Keast, 1959).

These facts, unless any proof to the contrary is obtained, show that it probably was the anticyclonic dry belt that varied in width during these climatic oscillations. They also show that, during the period in question, Australia did not appreciably change its geographical position.

In northern Australia the intensified circulation and the emergence of a large land area would cause a climate much more similar to that of South America today. Intertropical convergence and convectional thunderstorms

would bring torrential rains, separated by hot and steamy intervals. In the summer the inflow of equatorial air masses from the north was responsible for more humid climatic conditions. These air masses would lose some of their moisture before reaching Central Australia; this is why there could not be rain forests everywhere. In addition, no equatorial rain forest develops on highlands, so that the highland habitats of the Macdonnell Ranges and the Eastern Highlands must have been cooler and less luxuriant. South of the warm rainforest belt there was probably a belt of tropical open forest and savanna, the former along the margin of the rain forest and on the southeastern (windward) side of the continent, the latter on the western (leeward) side. There may have been a few smaller subhumid (woodland, savanna) areas in sheltered regions, as there now are in northeastern Brasil. These climatic conditions favoured gigantism among plants and animals, and the fossil record shows it: the giant grass-eating marsupials and flightless birds most likely developed in the more open country to the south of the equatorial rain forest, but fossils have been found in many areas.

From the dentition of some of the Pleistocene marsupials it has been inferred that they browsed on succulent reedlike plants (David and Browne, 1950). This would suggest a habitat somewhat like that of the modern hippopotamus.

So far *Thylacoleo carnifex* has not been recorded from Tasmania. *Diprotodon optatum* has been recorded from the mainland and from King Island. The giant Wombat *Phascolonus gigas* had very wide distribution. The giant Kangaroo *Palorchestes azael* has been recorded from eastern Australia and Tasmania. Species of the related *Sthenurus* have left remains in eastern Australia, King Island, and Western Australia. Most of the fossil Marsupial Wolves (Thylacines) were bigger than the living species (David and Browne, 1950).

Of the giant birds, *Dromornis*, almost as large as the largest moa, left fossil remains in Queensland and South Australia; the smaller *Genyornis* is recorded from the Paroo River, Q. and from such diverse South Australian localities as Lake Callabonna, Normanville, and Mount Gambier. Giant species of emu (*Dromaeus*) were also there.

The Riss/Würm interglacial. Beginning about 150,000 (Zeuner, Pfannenstiel) to 225,000 (Flint) years ago, the Riss/Würm interglacial lasted about 80,000 years and gave rise to the Monastirian sets of terraces in Europe. The Main Monastirian terraces (60 feet) were correlated by Crocker and Cotton (1946) with the Reedy Creek Range (70 to 75 feet) and with the West Avenue Range (85 to 90 feet) in South Australia. Tindale (1947) correlates them only with Reedy Creek.

David and Browne (1950) mentioned a submergence of 40 to 50 feet for this time and, in fact, Davies (1960) states that a terrace level of 40 to 50 feet is universally present on the Tasmanian coast, while Jennings (1958) finds it on King Island, where old dunes are also found. In Tasmania Davies (1960) also finds well-marked terraces at 60 to 75 feet.

Keble (1947) accepts the "Yolande/Margaret" interglacial as the Australian counterpart of Riss/Würm and ascribes to it the evidence of an arid climate noticed in deposits near Camperdown, V.

The Late Monastirian 25-foot terraces of Europe are correlated by Crocker

and Cotton (1946) and by Tindale (1947) with the Woakwine Range of South Australia (20 to 25 feet). The higher ones of the 12- to 22-foot terraces found by Davies (1960) in Tasmania almost certainly also belong to this phase. Radiocarbon dating of a 25-foot terrace at Port Fairy, V., gave an age of more than 35,000 years (Gill, 1955*b*). Jennings (1959) found marine terraces at 20 to 30 feet in King Island, and one sample of foraminifera from 15- to 30-foot beds, mostly similar to the present fauna, contained also two species from slightly warmer waters.

In Western Australia, shell beds at 22 feet above sea level along the Swan River were discussed by Somerville (1920) and a series of terraces formed by the Swan and Helena Rivers and culminating at 22 feet were described by Aurousseau and Budge (1921).

The Würm glacials. Now recognized as a complex of three glacials separated by two minor interglacials, the Würm glacials of Europe were correlated by Tindale (1947) with the "Margaret" Glacial of Tasmania, but this now needs complete revision. Zeuner places the beginning of the earliest phase at 115,000 years ago, that of the last phase at 23,000 years ago, Pfannenstiel at 110,000 and 18,000 respectively.

Jennings (1959) found a deltaic deposit containing large timber, abraded and not *in situ*, near sea level on King Island. Radiocarbon dating sets the timber at 37,500 ± 1900 years B.P. (Jennings *in litt.*). The timber was recognized as *Eucalyptus*, *Nothofagus*, and another kind of unidentified hardwood. Fossil pollen was overwhelmingly from celery-top pine (*Phyllocladus asplenii-folius*) but mountain pepper (*Drimys lanceolata*), myrtle beech (*Nothofagus cunninghamii*), and other dicotyledonous plants were represented; many fern spores including *Cyathea*, were recorded (Jennings, 1959).

Since *Drimys* and *Nothofagus* grow at higher levels in Tasmania, and since *Nothofagus* still grows in the Otways in southern Victoria, it may be inferred that the climate was slightly cooler than at present.

Genera of Proteaceae and Epacridaceae endemic to Tasmania "are always associated with cool habitats," while there are no Tasmanian endemic genera of Myrtaceae or Leguminosae. Burbidge (1960) suggests the "partial destruction of an early 'Australian type' vegetation during the Pleistocene glaciation and that only certain suitably adapted taxa survived." Because of the eustatic regression contemporary with glaciation this vegetation "was probably in communication with similar plant associations in southern Victoria across the exposed land of Bass Strait so that there has been an inadequate period for the development of Tasmanian endemic general. . . ."

The paleogeographical conditions under which genera from the temperate climates of the northern hemisphere reached northeast Queensland are completely unknown.

Gill (1955*b*) quotes alternate deposition of peat and marl in northwestern Tasmania radiocarbon-dated at 13,520 ± 540 B.P.

Keble (1947) considers that at Altona, V., there was a stage of emergence (W II?), then one of submergence (W II/III?) with deposition of shell beds and, later on, one of emergence (W III?) with formation of sand dunes. The same author states that deposits near Camperdown, V., show that part, at least, of the Margaret glacial was cool and rainy.

David and Browne (1950) list one Würm recession with a likely duration of about 16,000 years.

Tasmanian glaciation was studied for many years by Lewis (his work is best summarized in Lewis, 1934*a*; 1945) who believed there had been three distinct glacial stages: Malanna (icecap); Yolande (valley glaciers perhaps subdivided into two phases, Yolande I and II); Margaret (cirques).

It seemed unlikely to many that three successive glacial phases should so neatly differ in the nature of their physiographical work. I also thought that if the lowest snow line on Mount Kosciusko had not descended below 6000 feet (Browne, 1945, 1952), it was most unlikely that it should have been at 2000 feet, that is, 4000 feet lower, in Tasmania, only 5° 30′ farther south. Some authors had found difficulty in placing Lewis' scheme into the global pattern, and Lewis himself was at times very doubtful of the chronology (Lewis, 1934*a*). On the other hand Tindale (1947) very neatly took Malanna = Mindel, Yolande = Riss, Margaret = Würm. Flint (1947, 1957) preferred to regard Malanna = Riss, and Yolande and Margaret = Würm (two phases).

After field work and with the aid of aerial photography and aerophotogrammetry, Jennings and Ahmad (1957) tentatively believe that the ice-cap glaciation of western Tasmania was no older than Würm. The many hundreds of lakes are remarkably "new" and, for this reason, these authors think that their origin should not date back more than 30,000 or 40,000 years. Any traces of earlier glaciations have been obliterated. Glaciers and cirques in the area examined probably derive from the normal processes at the edge of the same icecap, which was much smaller than previously believed; Lewis' scheme of things becomes untenable (Jennings and Banks, 1958).

Not only was the Pleistocene snow line higher than Lewis had thought, but the ice-covered area and the area affected by glaciation generally were much smaller (Jennings and Banks, 1958) and the probable thickness of the ice much less (Jennings and Ahmad, 1957).

In northeast Tasmania only Ben Lomond (5160 feet) shows signs of having been glaciated (Nye and Lewis, 1927).

On Mount Kosciusko (Browne, 1945, 1952) the valley glaciers occupied valleys already formed by river erosion, at least some of which appear to have come into being in the interval between first and second glaciations. Erosion by valley glaciers was noticed at 4750 feet above sea level.

Browne (1952) maintains that "the valley-glaciers were vigorous and well-nourished, a condition which suggests that they were not remnants of a dwindling ice-sheet but belonged to a later and independent glaciation."

For similar reasons Browne prefers to separate the second from the third glacial phase. Jennings (*in litt.*) doubts whether any traces of earlier glaciations can still be seen. After further work, Browne and Vallance (1957) tentatively recognize at least three phases of valley-glaciation. The third glaciation was much less intense than the others and its effects seem to be confined to the higher altitudes. The cirque floors lie between 5950 and 6250 feet.

The Würm glacial may have favored the migration of the freshwater mussel *Lortiella*, which came to northwestern rivers from Asia, probably when the

Sahul Shelf was fully exposed and the distance from Asia much smaller. Two geographically distinct species are recognized (McMichael and Hiscock, 1958).

The study of Amphibia also discloses biogeographical peculiarities that can be explained best from paleogeographic evidence. Moore (1954) suggests that the presence in Tasmania of two very closely related species of frogs (*Crinia tasmaniensis* and *C. signifera*) may have been the result of a double invasion. The drying up of Bass Strait during two separate glacial phases would be enough to account for the two invasions and for the intervening specific differentiation.

Voous and van Marle (1949) discuss the distribution of the cuckoo shrikes, *Coracina*. They assume a Pliocene diffusion of *C. novaehollandiae* from southeast Asia, probably through Timor (Mayr, 1944), to Australia. "Speciation phenomena in the tropical Sunda-lowlands during the alternative Pleistocene periods of low and high sea level, have caused the development of a distinct lowland group (*striata*) from an ancestral *novaehollandiae* stem."

Voous and van Marle also think that "a backwardly directed wave has originated from the novaehollandiae-stock, radiating from eastern Australia. . . . The slightly developed subspecific distinctions in the most widely-spread species of this group (*C. papuensis*) indicate its rather recent dispersal through Cape York, New Guinea, Moluccas, and Celebes." They believe that "Pleistocene periods of low sea level connecting New Guinea with Australia and the Aru Islands, and bringing some of the Papuan Islands nearer to New Guinea, must have facilitated the dispersal of *C. papuensis* as a lowland species over the Sahul Land. Since its racial development is rather slight and only quantitative, and the same race of *C. papuensis* has been found in northwest Australia and the Aru Islands (*hypoleuca*; cf. Mayr, 1941), its colonization of northern Moluccas must have taken place rather recently, in the Pleistocene and Holocene." An outstanding number of plant genera are restricted to northeast Queensland and are probably recent immigrants; the Würm marine regression may be responsible for their arrival, although transport by ocean currents and by tropical cyclones can account for the introduction of many of them (Burbidge, 1960).

During the last Würm recession, with rising temperatures but still cooler than the present time, the Tasmanian and Kosciuskan glaciers finally vanished. Bass and Torres straits were flooded once more.

Torres Strait probably determined subspeciation in the freshwater mussel *Alathyria*, which has a fragmentary distribution that in part may actually be a complex of relict areas. McMichael and Hiscock (1958) recall the Eocene fossil *A. tamarensis* from northern Tasmania. Speciation and subspeciation have occurred repeatedly, and *A. pertexta* now occurs in three geographically and subspecifically separate populations; the drowning of Torres Strait may have isolated *A.p. magnifica* from *A.p. wardi.*

Polystadial Distributions

The biogeographical distributions that we know at present represent only the latest phase in a polystadial sequence. In our chronological review of

climatic changes and of their evidence we have endeavored to allocate each successive biogeographical phase to its paleoclimatic stage.

There are, however, some sequences that would become almost unintelligible if they were broken down to single phases.

Evidence for a previous east-west climatic continuity is found in the distribution of Galaxiidae fishes (minnow trouts), most of which live in fresh water, but a few are tolerant of brackish water (Stokell, 1953). Whitley (1947) lists the native minnow (*Galaxias occidentalis*) and the western mountain trout (*G. truttaceus hesperius*) as characteristic of the Vlaminghian fluvifaunula (fresh waters of the South-West). To these must be added the black-striped minnow (*Brachygalaxias pusillus nigrostriatus*, Shipway, 1953).

The native minnow (or western jollytail; Munro, 1957) occurs from Perth to Albany, and its closest relative, the common jollytail or eel gudgeon (*G. attenuatus*), is found in the coastal streams from southern Queensland to Tasmania and South Australia. "Apparently eggs are deposited during spring tides on rushes along tidal creeks. Larvae are released at succeeding spring tide, drift to sea and return upstream to fresh water" (Munro, 1957). Very similar to the latter species are the Tasmanian mud trout, *Galaxias upcheri*, which lives in tea-tree swamps and survives in thick mud during the dry season; the flat-headed minnow, *G. planiceps*, of the fresh waters of the Murray-Darling river system; the pieman or Mersey jollytail, *G. weedoni*, of the fresh waters of the Pieman and Mersey rivers, Lake Lawra, Tasmania; the lake trout, *G. parkeri*, of the Great Lake and nearby tarns (Munro, 1957). These are only a few examples of geographical speciation within this large genus.

Most significant is the geographical distribution of *Galaxias truttaceus*, the mountain trout. The type, *G.t. truttaceus*, is found in "coastal streams from brackish water to high altitudes" in south Australia, Victoria, the islands of Bass Strait, and Tasmania. The ocellated mountain trout, *G.t. scopus*, is found only in brackish water on Clarke Island, Bass Strait. The western mountain Trout, *G.t. hesperius*, is found only in fresh water near Albany, W.A.

The dwarf minnow, *Brachygalaxias pusillus*, lives in fresh water in Cardinia Creek, Victoria, where it was described in 1936. Its western counterpart, *B.p. nigrostriatus*, was discovered by Shipway near Marbellup Creek (Albany) in 1949. These subspecies obviously are the remnants of formerly continuous distributions. The continuity was possible only while fresh (or fresh and brackish) waters were found at very short intervals along the south coast.

To give some regular runoff on the shores of the Great Australian Bight, the annual rainfall would have to be about 30 inches greater than under the present conditions if its seasonal incidence were relatively uniform, and heavier still if there remained a prolonged seasonal drought.

The interpretation suggested here is that such conditions occurred in an early Pleistocene glacial, when the *G. attenuatus-occidentalis* superspecies had a continuous distribution. Later on, interglacial aridity excluded it from the region of the Great Australian Bight, and the uplift of the eastern highlands subdivided its eastern areale.

At a later stage a pluvial phase returned, and *G. truttaceus* and *B. pusillus* expanded westwards, reaching the Albany area. This expansion did not take

place very long ago (Würm?) if one may judge from the strong affinity of the subspecies concerned. Later still this continuity was broken once more by the southward spread of acidity (early Recent?) and what probably were the two ends of an extensive cline became separated and evolved slightly to become the subspecies known at present.

It should be pointed out that the only other species of *Brachygalaxias* is found in Chile; if this species is very closely related to the Australian ones and not simply a case of convergence, the most extraordinary biogeographical events must have occurred at some time in the past, unless we admit that ova of fresh-water species may travel for thousands of miles and last at least three months in sea water: in which case the whole distribution of Galaxiidae would not necessarily imply any climatic change.

Main *et al.* (1958) write that it seems plausible to account for the species of frogs present in southwestern Australia "by postulating an analogous series of repeated invasions from the primary speciation area in southeastern Australia. . . . These repeated invasions took place as the climates changed throughout the Pleistocene. In favourable times both Eyrean and Bassian forms invaded the southwest. In times of marked seasonal rainfall, however, not only was the west isolated from the east by a desert but in both *Crinia* and *Neobatrachus* all traces of earlier Bassian invaders were erased, so that at the present time only the most recent Bassian invader occurs (Littlejohn, 1959 considers one *Crinia* species as penultimate in the sequence). On the other hand, in *Heleioporus* both the last Bassian (*australiacus*) and the penultimate invader (*inornatus*) survive. However, in the Eyrean species earlier invasions were not erased by climatic deterioration and in each genus there are three populations, now species, each representing an earlier invasion."

The Bassian region would have been cool and with a "rather uniform rainfall," and the Eyrean region hotter and with a "more marked seasonal rainfall."

According to Main *et al.*, "the east-west connection with relatively uniform climatic conditions would be open for Bassian forms," but the same connection would be broken "as soon as contrasting seasonal conditions arose."

The distribution of Bassian forms may thus be explained very satisfactorily. Eyrean forms, however, present a more complicated problem. Main *et al.* consider it "probable that when the Bassian corridor was open along the south coast the inner fringe was subject to markedly seasonal and irregular rainfall at lower temperatures than those to which the earlier Western Eyrean invaders had become adapted." Thus some eastern Eyrean forms could migrate westward, but no western Eyrean form was able to migrate eastward.

These authors assume that during Pleistocene time three pluvials, separated by interpluvials, allowed three migrations of the faunal elements concerned. The Eyrean migratory elements persisted to the present time, but the earlier Bassian migrants to the west became extinct "during the seasonally arid interpluvials." Littlejohn (1959) thinks that, if *Crinia glauerti* is an earlier invader, the last pluvial phase must not have been "moist enough to enable a small species associated with permanent water to reach the southwest."

We owe to Serventy (1951) an analysis of the paleogeographical factors in

the speciation of the Chestnut-shouldered Wrens, *Malurus*. "At the present time *M. lamberti* is the most widely distributed species, and it may, perhaps, be closest to or even identical with the parental type, a member of the Eyrean fauna, with no surviving Bassian equivalent. The advent of an arid cycle in the climate must have cut off an isolated population in southwest Australia which gave rise to *M. elegans*, and another in the northeast (or northwest) part of the range, which gave rise to *M. amabilis*. The main bulk of *M. lamberti* survived as such in a refuge in southeastern South Australia. . . ."

We must endeavor to view these biogeographical changes in their correct time sequence. Too often now the Great Aridity of the early Recent (6000 to 4000 B.P.) is invoked as the *deus ex machina* by writers who, prior to the paper by Crocker and Wood (1947), had not even suspected its occurrence, not to mention its magnitude. Willett (1949) writes that probably the Climatic Optimum (= Arid Period of Australia) was a close approach to a true interglacial climate. "It appears that within a period of 5000 years in postglacial time world climate suffered an oscillation which probably represents half of that required to produce a Pleistocene interglacial-glacial cycle over a period of a hundred thousand years or more."

A re-evaluation of Pleistocene climates is needed; more than one "Great Aridity" certainly occurred in some part of Australia during the interglacials, and there is no need to crowd every biogeographical change into the 2000 years or so of the Recent Great Aridity, catastrophic as it may have been because of its suddenness.

Climates of the Recent Period

The Recent period is made to begin at about 10,000 B.P. A large fossil *Eucalyptus* at Melbourne was radiocarbon-dated at 8780 ± 200 B.P.; a bog moss now living above 4000 feet was associated with the tree and provides evidence of a colder climate confirmed by marine fossils. Pollen analyses and old river levels 12 to 15 feet higher than now are evidence of wetter conditions as well. Charcoal found in a river terrace was dated at 8500 ± 250 B.P. (Gill, 1955*b*).

The change by 8000 B.P. The climate was then warm and dry. Keble (1947) mentions the presence of yellow clay of windblown sand and tuff near Camperdown, V., as evidence of decreasing rainfall. At Altona, V., there is evidence of a rising sea level, of progressive submergence, and of the leveling of dunes. Keble (1947) considers that this is the time of the last drowning of Torres Strait and the beginning of the present separation of New Guinea from Australia. Europe at the time was going through its Boreal phase.

The trend by 7000 B.P. According to Rubinstein (1956) the general trend of climate in the northern continents by 7000 B.P. was warmer and more humid than in the previous millennium. Churchill (1960) discusses the finding on Rottnest Island, W.A., of fossilized blackboy (*Xanthorrhoea* sp.) radiocarbon-dated from 7090 ± 115 B.P. Keble (1947) discusses the rise of sea level and the progressive submergence and leveling of dunes at Altona, V. At this time there must have been a continuous belt of humid climate along the south coast. The same author places at 6000 B.P. the beginning of the period of

extreme aridity in Australia, probably corresponding to the climatic optimum of the Northern Hemisphere, which Willett (1953) believes to have been simultaneous in both hemispheres.

The Great Aridity: 5000 B.P. About 5000 B.P. there occurs a meridional expansion of the anticyclonic belt. Except at a few exposed localities, this time is one of heat and aridity, with general desiccation (David and Browne, 1950). In southwestern Victoria the prevailing winds, as evidenced by fossil layers of dust, blow from the northwest, not from the southwest as now (Gill, 1955*b*). The eruption of Mount Gambier volcano is later than 4710 ± 70 B.P., as shown by charcoal from the A horizon of soil immediately under the volcanic layer.

Churchill (1960) quotes evidence of peat deposition in fresh-water swamps on Rottnest Island, W.A., about 5000 B.P., with *Juncus* very plentiful. This is the time of high sea level (10-foot terrace) of the Flandrian transgression (Gill, 1955*b*). The sea may reach up to 15 or 20 feet above present sea level near Altona, V. (Keble, 1947), and invade the lower reaches of the rivers, as evidenced by fossil fishes and by fossil wood damaged by marine borers, 7 miles up the Maribyrnong River, C., dated 4820 ± 200 (Gill, 1955*b*). The mollusc *Anadara trapezia*, now uncommon near Perth and near Melbourne, was then extremely plentiful. Mangrove (*Avicennia officinalis*) was growing as far south as Port Noarlunga, S.A. (Cotton, 1960).

Fairbridge (1950, 1950*a*, 1956) gave detailed descriptions of the extensive 10-foot terrace now designated as the Peron Platform, from Point Peron south to Perth, W.A. On the Swan River, the 10-foot beach mentioned by Somerville (1920) is the site of the University campus.

Carrigy and Carrigy (1952) studied fossil shore features at Cottesloe, W.A., and recognized a former sea level of about 10 feet above present datum, "correlated with the warm climatic period of 4000 to 5000 years ago."

The fossil shells found belong to species living in the area at present and therefore do not provide any evidence of different climatic conditions from the present time (Kendrick, 1960); however there is evidence of a contemporary expansion of the "warmer" fauna (Cotton, 1960).

At the time of the Thermal Maximum a small lake near Camperdown, V., studied by Gill (1955*a*) was dried up, the prevailing wind was northwest, and probably "during the summers . . . great clouds of dust blew across the countryside in southeastern Australia. During the winters, however, it is likely that water gathered in the lake basins."

During the Pleistocene pluvials the increased availability of food had led to a great increase in the numbers of marsupials, especially *Diprotodon.*

The spread of aridity led these animals to concentrate around the many lakes, still fresh at the time, where water and feed could still be found. In this way many individuals became bogged in Lake Callabonna (north of Lake Frome, S.A.) where skeletons and individual bones were recovered at the end of the last century. Fossil stomach contents included fragments of Chenopodiaceae and Ameranthaceae or Nyctaginaceae, and it may be inferred that at the arid time when *Diprotodon* was dying out, the vegetation may have been similar to that of today's Saltbush-Bluebush communities (Burbidge, 1960).

The southward spread of aridity reached the Great Australian Bight, thus dividing the southwestern from the southeastern forests. The rapid southward advance of the desert enabled only relatively few plant species to survive in the extreme southwest, very few on the Flinders and Mount Lofty ranges, and many on the Eastern Highlands and in Tasmania. Isolation of all these remnants was a powerful factor in bringing forth differentiation. When climatic conditions later on permitted the dispersal of the refugees, many new forms had appeared (Crocker and Wood, 1947). Serventy and Whittell (1948) discuss the geographical distribution of western birds. They quote as examples 19 species which, conditioned to a dry environment or remarkably tolerant of it, extend across the continent as far as the "Great Dividing Range," but do not usually cross into the rainier region to the east. The name Eyrean is usually given to this dry areale, and the name Bassian to the eastern and wetter areale.

These authors point out that the southwest has many species of birds whose affinities and relationships are with Bassian (southeastern) species. They quote only three species that range uninterruptedly across the southern part of the continent; their areale includes a narrow corridor on or around the Nullarbor Plain.

Another 19 bird species are cited as examples of groups that have a discontinuous areale; they occur both in the southwest and in the southeast of the continent, but are divided by an enormous gap in correspondence of the Great Australian Bight. These are species that were divided relatively recently, and/or were less prone to evolutionary changes.

Another seven southwestern bird species are mentioned that have seven corresponding and closely related species in the Bassian region. These pairs of species may have been separated earlier, and/or may have been more prone to evolutionary change.

According to the same authors, at least 18 southeastern species are entirely absent from the southwestern area, where the environment is still suitable for many of them. Were they ever represented there in the past, only to become extinct later?

From this ornithological evidence the fact emerges that 26 species or superspecies have similar habitats and a very similarly discontinuous areale. The discontinuity can be accounted for only by the arising of an unfavorable environment north of the Great Australian Bight.

Serventy (1951) thinks that the chestnut-shouldered wrens were already differentiated into the species *Malurus pulcherrimus*, *M. lamberti*, *M. elegans* and, in northern Australia, *M. amabilis*. The onset of aridity caused notable changes. "The ancestors of *M. pulcherrimus* must have been left behind in southwestern Australia by the retreating principal population of *M. lamberti*, where they probably narrowly compressed *M. elegans* into the inner fastnesses of the south-west humidity refuge. In the north *M. amabilis* probably became divided up into . . . *M.a. amabilis* in northeastern Queensland and *M.a. dulcis* in the Northern Territory and the Kimberley Division of Western Australia."

More recently Serventy (1953) discusses the various pairs of Eyrean and Bassian bird forms, giving a reconstruction of the paleogeographical position

of *Pardalotus*, *Myzantha*, and other genera. Recent climatic conditions are mentioned.

The presence of fossilized bones of the koala (*Phascolarctos cinereus*) in the southwestern caves (Glauert, 1948) gives a definite indication of the raininess of earlier phases. The koala is strictly arboreal; it feeds on shoots of a few species of eucalyptus. Differently from other arboreal marsupials, the koala is very slow in its movements and, once on the ground, could not flee from any predators. Hence the life of the species depends on the availability of suitable food trees, close enough to enable the individuals to escape readily from ground predators. In Western Australia at least three species of eucalyptus are palatable to the koala: York gum (*E. loxophleba*), Wandoo (*E. redunca*), Flooded gum (*E. rudis*). The York gum is a tree of the open woodland, growing much farther east than the area where the fossil remains were found, under a mean annual rainfall lower than 20 inches. The Wandoo covers woodland country between the southwestern forests (of Jarrah, *E. marginata*) and the York gum woodland, with a mean annual rainfall of 20 to 25 inches. It also grows throughout the rainier forest area farther west, but there it is restricted to the red loam on dykes of dolerite. Thus neither the York gum nor the Wandoo is likely to have supported a population of koalas under climatic conditions similar to the present ones.

The flooded gum, on the other hand, is now restricted to the banks of water courses. Obviously the present climate is too dry for its optimum growth. The rainfall of its areale is 40 to 50 inches a year. In order to grow away from water courses it would need either a greatly increased rainfall, probably exceeding 50 to 60 inches a year if the present summer drought were still present, or a rainfall not greater than the present one, but with no summer drought.

Another hypothesis may be advanced. Jarrah (*E. marginata*) forms great forests under the present annual rainfall of 40 to 50 inches. Jarrah is a strange tree, because it takes several years, as an overage seedling, to establish a lignotuber, without which it cannot grow to tree size. This lignotuber enables the tree to overcome the long seasonal drought, made worse by the dryness of the lateritic duricrust on which Jarrah grows. Wandoo cannot grow in such a harsh environment. However, should there be no summer drought, Wandoo would soon outgrow Jarrah on the better soils and form true forests there. Whether it would also oust Jarrah from the lateritic country cannot be surmised. This alternative hypothesis also requires a climate with uniform rainfall.

The southwestern forests are more than 1400 miles away from the nearest southeastern forests. In order to bring about the climatic conditions needed for a continuous, albeit very narrow, belt of woodland right across southern Australia, the annual rainfall of the Nullarbor plain would have to be 20 to 30 inches higher than at present (Gentilli, 1951). Without a continuous belt of woodland the highly specialized koala could not have populated these widely separated forest areas.

Thus the past distribution of the koala presupposes a period without or nearly without a summer drought and a (not necessarily concomitant, but not later) period with an annual rainfall 20 to 30 inches higher than at present.

The finding of bones of Tasmanian Devil (*Sarcophilus harrisii*) and Tasmanian Wolf (*Thylacinus cynocephalus*) in the same caves (Glauert, 1948) sets a similar problem: how could these not very mobile carnivores bridge the gap between eastern and western Australia? Under present climatic conditions and prior to the rabbit invasion these carnivores could not have lived in the Nullarbor plain, where only small marsupials existed (Lundelius, 1957). Enough food to support them across the plain could be available only under much wetter climatic conditions, wet enough to wash the cyclic salt out of the ground and wet enough to make possible the growth of a much denser vegetation.

The fact that the fossil deposits listed by Glauert (1948) include bones of very big marsupials which are now extinct (the gigantic wombatlike *Nototherium mitchelli*; *Palorchestes sp.* = giant kangaroo; *Thylacoleo carnifex* = marsupial lion; *Zaglossus hacketti* = giant echidna) may further support the contention that a more luxuriant vegetation was formerly available, and that these arid conditions very soon caused the extinction of these marsupials in the southwest.

It may well be at this time that the drying up of the northwestern rivers divided the freshwater mussel *Westralunio* into a northern population (*W. flyensis* and *W. albertisi*) and a southwestern population (*W. carteri*).

All the evidence points to a hot and very dry climate. European authors write of a "Climatic Optimum," a term that well applies to a continent just emerged from glacial conditions, but Australian writers (for example, Crocker and Wood, 1947) have referred to the "Arid Period." Gill (1955*a*) prefers the expression thermal maximum.

This may be the time of formation of the very extensive sand ridges that now cover about 523,000 square miles (Browne, 1945). The climate of today is arid over a great part of the Australian interior, but not arid enough to lead to the formation of such ridges which, to the contrary, now carry a sparse vegetation and are fixed.

It seems likely that this thermal maximum caused the formation of the later parna (wind-borne eroded soil) and the deposition of dust near Camperdown, V.

About 4000 B.P. By about 4000 B.P. *Juncus* had nearly disappeared from Rottnest Island, W.A., and pollen from Cyperaceae had become dominant. Fossilized *Callitris* wood from Rottnest Island, W.A., was dated 3950 ± 130 B.P. (Churchill, 1960). This may be the onset of cooler and moister conditions (David and Browne, 1950). The shore progressively emerges, and Keble (1947) finds uncovered dunes with a leveled surface near Altona, V. Black clay occurring with windblown sand and tuff near Camperdown, V., may be a sign of increasing rainfall (Keble, 1947), leading to a climate wetter than the present (Gill, 1955*a*).

The area of mangrove growth in South Australia is gravely restricted, and part of the warm-water fauna is wiped out, "probably by climatic change, and later silting by fine sand" (Cotton, 1960).

This is the time when a slow deterioration begins in the climate of the Northern Hemisphere (Willett, 1949).

A pluvial phase: 3000 B.P. A wetter climate than at present is shown by higher levels in lakes and rivers of Victoria about 3000 B.P. Charcoal from the Keilor Terrace near Melbourne is dated 3100 ± 160 (Gill, 1955*b*). Formation of peat occurs. This is the stormy (sub-Atlantic) phase of Western European climate, which is followed by a sudden and great climatic deterioration, the "Little Ice Age" of northern authors. Willett (1949) points out that there was a lag of 4 centuries in this phase between Europe and North America, and it may well be that Australian chronology also lags, probably because of differences in the size and geographical position of the land mass.

In Australia the climate was sufficiently wet to cause the formation of laterite (Jessup, quoted by Burbidge, 1960). The revival of the north-western rivers may have enabled the Mussel *Velesunio angasi* to colonize the waters of Arnhem Land and the Kimberleys, thus creating the biological wedge that now even more effectively separates the two *Westralunio* populations.

After the earlier arid period "it is probable that the winter rainfall system" may have extended inland and northwards, allowing "a northward migration of southern elements. This may explain the presence, in central Australia, of such species as *Hakea multilineata* Meissn. and *Burtonia polyzyga* (F. Muell) Benth. in the MacDonnell Range and isolated occurrences of mallees such as *Eucalyptus oleosa* F. Muell" (Burbidge, 1960).

If a pluvial phase affects northern and southern Australia at the same time, we may expect a southward advance of the rain-forest communities from the very limited refuges in the Cairns-Innisfail area.

Burbidge (1960) considers the discontinuities in rain-forest formation in New South Wales "more likely to have resulted from climatic changes since and not during the Pleistocene. The pluvial stage of the Pleistocene was followed by a severe arid period but since then there seems to have been at least two pluvial periods, separated and followed by arid periods the last of which is still in progress." While the climate was sufficiently humid and with no dry season the tropical rain forest could have extended southwards. Each arid stage would cause a northward retreat.

Harrison (1960) states that Australian tropical rain forests have a very low faunal diversity. I take this fact as an indication of the recent establishment of the tropical rain-forest communities now existing in Australia, and of the fact that the immigration of many floral and faunal elements from the north has been by sea. Plants migrate by sea far better than animals. The floristic-vegetational bridgehead is established, but the faunistic invaders are still in the process of coming. This theory agrees with the facts stressed by Tate (1953), that is, that only a few New Guinea species of mammals (11 ?) have entered Australia, and that their occurrence is very uneven, with notable gaps.

From about 2600 to 2100 B.P. the Dunkirkian submergence took place, as shown in Western Australia by the 5-foot terraces of the Abrolhos Islands.

About 2000 B.P. On Rottnest Island, W.A., by about 2000 B.P., pollens from Euphorbiaceae and from *Casuarina* sp. were equally plentiful, with pollen from Cyperaceae being third and on the decline. Pollens from *Eucalyptus* and from Compositae were present throughout the record but in

gradually decreasing amounts. *Melaleuca* appears in the last 3000 years (Churchill, 1960). The rainfall is somewhat greater (David and Browne, 1950). This minor pluvial stage corresponds to a slightly cooler and drier stage in the northern continents (Rubinstein, 1956).

From 1000 B.P. to historical records. On Rottnest Island pollen from Gramineae only appeared in the last 1500 years of the fossil record (Churchill, 1960). In Victoria there is evidence of the presence of the broad-toothed rat on the western coastal plain; this animal is now limited to the Otway Mountains. Charcoal from this area dates from 1177 ± 175 (Gill, 1955*b*).

The Tasmanian devil (*Sarcophilus*) was used for food by the aborigines of Victoria, where it is now extinct, as shown by charcoal from kitchen midden dated 538 ± 200 (Gill, 1955*b*).

Lundelius (1957) found in the superficial deposits of some caves of the Nullarbor area remains of mammals (*Bettongia penicillata*, *B. lesueuri*, *Pseudocheirus occidentalis*, *Phascogale calura*) known at present from the rainier southwestern areas. Special care was taken to collect only from the surface and down to a depth of one foot, in order to exclude older fossils. These remains show that the respective species lived in that region not very long ago. "The cause of the recent disappearance of the native mammals in this area in unknown. It could be due to climatic change, the effect of introduced mammals, or both." Competition from introduced mammals could explain the disappearance of the first three species, but not of the carnivorous *Phascogale*. Climatic change (aridity) seems a much more likely common factor.

It may be relevant to point out that even within the short period of observational records in Australia there have been noticeable fluctuations in the rainfall of the southwestern region (Gentilli, 1952). These changes may be evidence merely of the statistical variability of the rainfall at the present time but, on the other hand, should they continue over a longer period, their magnitude is sufficient to give rise to a definite climatic change. Observations of birds show some remarkable extensions of areale. For instance, dry-climate birds are slowly invading the wetter southwestern and southeastern parts of the continent (Serventy, private communication).

References

Albrecht, F. 1958. Untersuchungen über den Wärmeumsatz an der Meeresoberfläche und die Meeresstromungen im Indischen Ozean. Geof. Pura e Appl. **39:** 194–215.

Alissow, B. P., O. A. Drosdow & E. S. Rubinstein. 1956. Lehrbuch der Klimatologie. Berlin, Germany.

Antevs, E. 1928. The last glaciation. Am. Geogr. Soc. Research Ser. No. 17 (New York).

Arrhenius, G. 1950. Late cenozoic climatic changes as recorded by the equatorial current system. Tellus. **2:** 83–88.

Aurousseau, M. & E. A. Budge. 1921. The terraces of the Swan and Helena Rivers. J. & Proc. Roy. S. W. Australia. **7:** 24–43.

Baker, G. & E. D. Gill. 1957. Pleistocene emerged marine platform, Port Campbell, Victoria. Quatern. **4:** (abstr. 1–14).

Balme, B. E. & D. M. Churchill. 1959. Tertiary sediments at Coolgardie, Western Australia. J. Roy. Soc. W. Australian. **42:** 37–43.

Barber, H. N. 1954. Adaptive gene substitutions in Tasmanian eucalypts: I. Genes controlling the development of glaucousness. Evol. **9:** 1–14.

Barber, H. N. & W. D. Jackson. 1957. Natural selection in action in Eucalyptus. Nature. **179:** 1267–1269.

BEAVIS, F. C. 1959. Pleistocene glaciation on the Bogong high plains. Australian J. Sci. **21:** 192.

BLANC, A. C. 1942. Variazioni climatiche e oscillazioni della linea di riva nel Mediterraneo centrale durante l'Era glaciale. Geol. d. Meere u. Binnengew. **5:** 137–219.

BODEN, R. W. 1958. Differential frost resistance within one Eucalyptus species. Australian J. Sci. **21:** 84–86.

BOSWELL, P. G. H. 1940. Climates of the past: a review of the geological evidence. Quart. J. Roy. Meteorol. Soc. **66:** 249–274.

BROOKS, C. E. P. 1947. Unsolved problem of climatic change. Meteorol. Mag. **76:** 126–129, 147–151.

BROOKS, C. E. P. 1949. Climate Through the Ages. London, England.

BROOKS, C. E. P. 1951. Selective annotated bibliography on climatic changes. Meteorol. Abstr. Bibl. **1:** 446–475.

BROOKS, C. E. P. 1951*b*. Geological and historical aspects of climatic change. *In* Compendium of Meteorology. Boston, Mass.

BROWNE, W. R. 1945. An attempted post-Tertiary chronology for Australia. Proc. Linn. Soc. N.S.W. **70:** v–xxiv.

BROWNE, W. R. 1952. Pleistocene glaciation in the Kosciusko region. *In* Sir Douglas Mawson Anniversary Volume. Univ. Adelaide. Adelaide, Australia.

BROWNE, W. R. 1957. Pleistocene glaciation in the Commonwealth of Australia. J. Glaciol. **3:** 111–115.

BROWNE, W. R. & T. G. VALLANCE. 1957. Notes on some evidences of glaciation in the Kosciusko region. Proc. Linn. Soc. N.S.W. **82:** 125–144.

BURBIDGE, N. T. 1960. The phytogeography of the Australian region. Australian J. Botany. **8:** 75–212.

BUTLER, B. E. 1956. Parna—an Aeolian clay. Australian J. Sci. **18:** 145–151.

BUTLER, B. E. 1958. Depositional systems of the riverine plain of S.E. Australia in relation to soils. C.S.I.R.O. Soil Pub. 10.

BUTLER, B. E. 1959. Periodic phenomena in landscapes as a basis for soil studies. C.S.-I.R.O. Soil Pub. 14.

CAIN, A. J. 1955. A revision of *Trichoglossus haematodus* and of the Australian Platycercine parrots. *Ibid.* **97:** 432–479.

CARR, S. G. M. & A. B. COSTIN. 1955. Pleistocene glaciation in the Victorian Alps. Proc. Linn. Soc. N.S.W. **80:** 217–228.

CARRIGY, M. A. & S. CARRIGY. 1952. Evidence of a mid-recent change of sea-level at Cottesloe. W. Australian Naturalist. **3:** 147–151.

CHARLESWORTH, J. K. 1957. The Quaternary Era. London, England.

CHURCHILL, D. M. 1959. Late Quaternary eustatic changes in the Swan river district. J. Roy. Soc. W. Australia. **42:** 53–55.

CHURCHILL, D. M. 1960. Late Quaternary changes in the vegetation of Rottnest Island. W. Australian Naturalist. **7:** 160–166.

COOKSON, I. 1953. The identification of the sporomorph *Phyllocladidites* with *Dacrydium.* Australian J. Botany. **1:** 64–70.

COOKSON, I. 1954. The occurrence of an older tertiary microflora in western Australia. Australian J. Sci. **16:** 37–38.

COOKSON, I. & S. L. DUIGAN. 1950. Fossil Banksiae from Yallourn, Victoria, with some notes on the morphology and anatomy of living species. Australian J. Sci. Research. **B3:** 133–164.

COOKSON, I. & K. M. PIKE. 1953. The Tertiary occurrence and distribution of *Podocarpus* (section *Dacrycarpus)* in Australia and Tasmania. Australian J. Botany. **1:** 71–82.

COOKSON, I. & K. M. PIKE. 1954. Some dicotyledonous pollen types from Cainozoic deposits in the Australian region. Australian J. Botany. **2:** 197–218.

COSTIN, A. B. 1959. Vegetation of high mountains in Australia in relation to land use. *In* Keast *et al.* (1959).

COTTON, B. C. 1960. Recent alterations in range and abundance of marine invertebrates in South Australia. W. Australian Naturalist. **7:** 137–141.

COTTON, L. A. 1947. The pulse of the Pacific. J. Proc. Roy. Soc. N.S.W. **80:** 41–76.

CROCKER, R. L. 1946. Post-miocene climatic and geologic history. . . . C.S.I.R. Bull. **193**.

CROCKER, R. L. 1959. Past climatic fluctuations and their influence upon Australian vegetation. *In* Keast *et al.* (1959).

CROCKER, R. L. & B. C. COTTON. 1946. Some raised beaches of the lower southeast of South Australia and their significance. Trans. Roy. Soc. S. Australia. **70:** 64–82.

CROCKER, R. L. & J. G. WOOD. 1947. Some historical influences on the development of the South Australian vegetation communities. . . . Trans. Roy. Soc. S. Australia. **71:** 91–136.

Daly, R. A. 1934. The Changing World of the Ice Age. New York, N. Y.
David, T. W. E. 1923. Geological evidence of the antiquity of man in the Commonwealth. . . . Pap. Proc. Roy. Soc. Tasmania. **1923:** 109–150.
David, T. W. E. & W. R. Browne. 1950. The Geology of the Commonwealth of Australia. London, England.
Davies, J. L. 1960. Quaternary strandlines in Tasmania. *In* Summary of the Report of the A.N.Z.A.A.S. Sections C and P Committee. . . . Australia J. Sci. **23:** 75.
Dawson, E. W. 1960. Advances in the knowledge of the small subfossil birds of New Zealand, and their biogeographical significance. XV Intern. Congr. Zool. Sect. V. paper 22.
Downes, R. G. 1954. Cyclic Salt as a Dominant Factor in the Genesis of Soils in South-eastern Australia. Australian J. Agr. Research. **5:** 448–464.
Dozy, J. J. 1937. "Geologie" and "Topografie." *In* Naar de eeuwige sneeuw van tropisch Nederland. A. H. Colijn, Ed. Amsterdam, Holland.
Duff, R. 1949. Pyramid Valley (Christchurch).
Dulhunty, J. A. 1945. On glacial lakes in Kosciusko region. J. Roy. Soc. N.S.W. **79:** 143–152.
Ekman, L. 1953. Zoogeography of the sea. London, England.
Endean, R. 1957. The Biogeography of Queensland shallow-water Echinoderm Fauna (excluding Crinoidea), with a rearrangement of the faunistic provinces of tropical Australia. Australian J. Marine Freshwater Research. **8:** 233–273.
Fairbridge, R. W. 1950. Recent and Pleistocene coral reefs of Australia. J. Geol. **58:** 330–401.
Fairbridge, R. W. 1950*a*. The geology and geomorphology of Point Peron, Western Australia. J. Roy. Soc. W. Australia. **34:** 35–72.
Fairbridge, R. W. 1953. The Sahul Shelf, northern Australia. J. Roy. Soc. W. Australia. **37:** 1.
Fairbridge, R. W. 1956. Western Australia. *In* Australian and New Zealand Research in Eustasy. Australian J. Sci. **19:** 57–58.
Fairbridge, R. W. 1958. Dating the latest movements of the Quaternary sea level. Trans. N.Y. Acad. Sci. Ser. II. **20**(6)**:** 471–482.
Fleming, C. A. 1942. Past climatic changes and seabird speciation. Australia J. Sci. **4:** 113–117.
Fleming, C. A. 1953. Quaternary geochronology in New Zealand. Actes IV Congr. Int. Quat. (Roma and Pisa).
Fleming, C. A. 1957. Trans-Tasman relationships in natural history. *In* Science in New Zealand. Wellington, England.
Fleming, C. A. 1958. Darwinism in New Zealand: some examples, influences, and developments. Proc. Roy. Soc. New Zealand. **86:** 65–86.
Flint, R. F. 1947. Glacial Geology and the Pleistocene Epoch. New York, N.Y.
Flint, R. F. 1957. Glacial and Pleistocene Geology. New York, N.Y.
Flohn, H. 1952. Allgemeine atmosphärische Zirkulation und Paläoklimatologie. Geol. Rundsch. **40:** 153–178.
Gentilli, J. 1949. Foundations of Australian bird geography. Emu. **49:** 85–129.
Gentilli, J. 1951. Bioclimatic changes in Western Australia. W. Australia Naturalist. **2:** 175–184.
Gentilli, J. 1952. Present Climatic Fluctuations in Western Australia. W. Australian Naturalist. **3:** 155–165.
Gentilli, J. 1958. A Geography of Climate (Perth).
Gill, E. D. 1955*a*. The Australian "arid period". Australian J. Sci. **17:** 204–206.
Gill, E. D. 1955*b*. Radiocarbon Dates for Australian Archaeological and Geological Samples. Australian J. Sci. **18:** 49–52.
Gill, E. D. 1956. Radiocarbon dating for glacial varves in Tasmania. Australian J. Sci. **19:** 80.
Gill, E. D. 1957. The Pliocene-Pleistocene boundary in Australia. Australian J. Sci. **20:** 56–57.
Gill, E. D. 1959. Penguins—past and present. Victoria Natur. March. **1959**.
Glaessner, M. F., B. McGowran & M. Wade. 1960. Discovery of a Kangaroo-Bone in the Middle Miocene of Victoria. Australian J. Sci. **22:** 484–485.
Glauert, L. 1934. The distribution of marsupials in Western Australian. J. Roy. Soc. W. Australia. **19:** 17–32.
Glauert, L. 1948. The cave fossils of the South-West. W. Australian Naturalist. **1:** 100–104.
Godwin, H., R. P. Suggate & E. H. Willis. 1958. Radiocarbon dating of the eustatic rise in ocean level. Nature. **11:** 1518–1519.

GOOD, R. 1957. Some problems of Southern floras with special reference to Australia. Australian J. Sci. **20:** 41–44.

GOOD, R. 1958. The Biogeography of Australia. Nature. **181:** 1763–1765.

GRAHMANN, R. 1952. Das Eiszeitalter und der Übergang zur Gegenwart (Remagen).

HARRISON, J. L. 1960. Faunal diversity in Australian tropical rain forest. Australian J. Sci. **22:** 424–426.

HARRISON, L. 1926. The composition and origins of the Australian Fauna, with special references to the Wegener hypóthesis. Proc. of A.N.Z.A.A.S. (Perth).

JENNINGS, J. N. 1957. Coastal Dune-Lakes as Exemplified from King Island, Tasmania. Geogr. J. **123:** 59–70.

JENNINGS, J. N. 1958. The coastal geomorphology of King Island, Bass Strait, in relation to changes in the relative level of land and sea. Records Queen Victoria Museum.

JENNINGS, J. N. & N. AHMAD. 1957. The Legacy of an Ice Cap. Australia Geogr. **7:** 62–75.

JENNINGS, J. N. & M. R. BANKS. 1958. The Pleistocene glacial history of Tasmania. J. Glaciol. **3:** 298–303.

KAY, G. R. 1931. Classification and duration of the Pleistocene period. Bull. Geol. Soc. Am. **42:** 425–466.

KEAST, A. 1959. Relict Animals and Plants of the Macdonnell Ranges. Austr. Mus. Mag. (**1959**): 81–86.

KEAST, A., R. L. CROCKER & C. S. CHRISTIAN (Eds). 1959. Biogeography and Ecology in Australia (den Haag).

KEBLE, R. A. 1947. Australian Quaternary climates and migration. Mem. Natl. Museum. (Vic.): **15:** 28–81.

KENDRICK, G. W. 1960. The Fossil Mollusca of the Peppermint Grove Limestone. . . . W. Australian Naturalist. **7:** 53–66.

KRAUS, E. B. 1954. Secular changes in the rainfall regime of SE. Australia. Quart. J. Roy. Meteorol. Soc. **80:** 591–601.

KRAUS, E. B. 1955. Secular changes of tropical rainfall regimes. Quart. J. Roy. Meteorol. Soc. **81:** 198–210.

KRAUS, E. B. 1960. Synoptic and dynamic aspects of climatic change. Quart. J. Roy. Meteorol. Soc. **86:** 1–15.

LAM, H. J. 1934. Materials towards a study of the flora of the islands of New Guinea. Blumea. **1:** 115–159.

LANDSBERG, H. 1949. Climatology of the Pleistocene. Bull. Geol. Soc. Am. **60:** 1437–1442.

LANGFORD-SMITH, T. 1960. The Dead River Systems of the Murrumbidgee. Geogr. Rev. **50:** 368–389.

LAUNCESTON, N. S. No. 11.

LEWIS, A. N. 1934*a*. A correlation of the Tasmanian Pleistocene glacial epochs and deposits. Pap. Proc. Roy. Soc. Tasmania. (**1933**): 67–76.

LEWIS, A. N. 1934*b*. A correlation of the Tasmanian Pleistocene Raised Beaches and River Terraces. Pap. Proc. Roy. Soc. Tasmania. (**1934**): 75–86.

LEWIS, A. N. 1945. Pleistocene Glaciation in Tasmania. Pap. Proc. Roy. Soc. Tasmània. (**1944**): 41–56.

LITTLEJOHN, M. J. 1959. Call differentiation in a complex of seven species of *Crinia* (Anura, Leptodactylidoe). Evol. **13:** 452–468.

LORENTZ, H. A. 1911. An expedition to the Snow Mountains of New Guinea. Geogr. J. **37:** 477–500.

LUCAS, A. H. S. & W. H. D. LE SOUEF. 1909. The Animals of Australia. Melbourne, Australia.

LUNDELIUS, E. 1957. Additions to knowledge of the ranges of Western Australian mammals. W. Australian Naturalist. **5:** 173–182.

MCMICHAEL, D. E. & I. D. HISCOCK. 1958. A monograph of the freshwater mussels (Mollusca:Pelecypoda) of the Australian region. Austr. J. Marine Freshwater Research. **9:** 372–508.

MAIN, A. R., A. K. LEE & M. J. LITTLEJOHN. 1958. Evolution in three genera of Australian frogs. Evol. **12:** 224–233.

MAIN, A. R., M. J. LITTLEJOHN & A. K. LEE. 1959. Ecology of Australian Frogs. Biogeography and Ecology in Australia. 396–411 (den Haag).

MAYR, E. 1941. List of New Guinea birds. Am. Museum Nat. History. New York, N.Y.

MAYR, E. 1944. Timor and the Colonization of Australia by Birds. Emu. **44:** 113–130.

MAYR, E. 1953. Fragments of a Papuan ornithogeography. Proc. 7th Pac. Sci. Congr. **4:** 11–19.

MAYR, E. 1954. Change of Genetic Environment and Evolution. *In* J. Huxley *et al.*, Ed. Evolution as a Process. London, England.

Moore, J. A. 1954. Geographic and genetic isolation in Australian amphibia. Am. Naturalist. **88:** 64.
Mueller, F. von. 1882. Report on the Forest Resources of the Colony. *In* General Information Respecting the Present Condition of the Forests etc. (Perth).
Munro, I. S. R. 1957. Handbook of Australian Fishes, suppl. to Fisheries Newsletter. **16**(2): 15–18.
Nicholls, G. E. 1926. The composition and biogeographical relations of the fauna of Western Australia. Rept. of A.N.Z.A.A.S. **21:** 93–138.
Nilsson, E. 1949. The Pluvials of East Africa. *In* Glaciers and Climate, Geogr. Ann. 1949.
Nye, P. B. & A. N. Lewis. 1927. Geology. *In* Handbook to Tasmania (Hobart).
Pfannenstiel, M. 1944. Die diluvialen Entwicklungsstadien und die Urgeschichte von Dardanellen, Marmarameer und Bosporus. Geol. Rundschau. **34:** 342–344.
Plass, G. N. 1956. The carbon dioxide theory of climatic change. Tellus. **8:** 140.
Pryor, L. D. 1959. Evolution in Eucalyptus. Australian J. Sci. **22:** 45–(49) ix.
Rathjens, C. 1954. Das Problem der Gliederung des Eiszeitalters in physischgeographischer Sicht. Münchn. Geogr. H. **6:** 1–68.
Reiner, E. 1960. The Glaciation of Mount Wilhelm, Australian New Guinea. Geogr. Rev. **50:** 490–503.
Richards, H. C. 1939. Recent sea-level changes in Eastern Australia. Proc. 6th Pacif. Sc. Congr. **:** 853–856.
Rubinstein, E. S. 1956. Klimaänderungen und Klimaschwankungen. *In* Alissow *et al.*
Serventy, D. L. 1951. The evolution of the chestnut shouldered wrens (*Malurus*). Emu. **51:** 113–120.
Serventy, D. L. 1953. Some speciation problems in Australian birds. Emu. **53:** 131–145.
Serventy, D. L. & H. M. Whittell. 1948. Birds of Western Australia (Perth).
Shapley, H., Ed. 1953. Climatic Change. Cambridge, Mass.
Shipway, B. 1953. Additional records of fishes occurring in the 'fresh' waters of western Australia. W. Australian Naturalist. **3:** 173–177.
Simpson, G. C. 1930. The climate during the Pleistocene period. Proc. Roy. Soc. Edinburgh. **50:** 262–296.
Simpson, G. C. 1938. Ice Ages. Smiths. Rept. **:** 289–302.
Somerville, J. L. 1920. Evidences of uplift in the neighborhood of Perth. J. & Proc. Roy. S. W. Australia. **6:** 5–20.
Stearns, H. T. 1945*a*. Late geological history of the Pacific basin. Am. J. Sci. **243:** 614–620.
Stearns, H. T. 1945*b*. Eustatic shore lines in the Pacific. Bull. Geol. Soc. Am. **56:** 1071–1078.
Stokell, G. 1953. The distribution of the family Galaxiidae. Proc. Sev. Pac. Sci. Congr. **4:** 48–52.
Süssmilch, C. A. 1941. The climate of Australia in past ages. J. Proc. Roy. Soc. N.S.W. **75:** 47–64.
Swain, E. H. F. 1928. The Timbers and Forest Products of Queensland. Brisbane, Australia.
Tate, G. H. H. 1953. Notes on mammalian distribution in the Cape York Peninsula. Proc. 7th Pac. Sci. Congr. **4:** 32–38.
Tate, R. 1889. On the influence of physiographic changes in the distribution of life in Australia. Rept. Australia Assoc. Advancement Sci. **1:** 312.
Taylor, T. G. 1919. Climatic Cycles and Evolution. Geogr. Rev. **8:** 292.
Taylor, T. G. 1920. Australian Meteorol. Oxford, England.
Taylor, T. G. 1927. Environment and Race. Oxford, England.
Taylor, T. G. 1937. Environment, Race, and Migration. Toronto, Ont., Canada.
Teichert, C. 1947. Stratigraphy of western Australia. J. Proc. Roy. Soc., N.S.W. **80:** 81–142.
Teichert, C. 1950. Late Quaternary changes of sea-level at Rottnest Island, Western Australia. Proc. Roy. Soc. Victoria. **59:** 63–79.
Tindale, N. B. 1947. Subdivision of Pleistocene time in South Australia. Record. S. Australia Museum. **8:** 619–652.
Trevisan, L. 1940. I limiti nivali attuali e würmiani in Italia. . . . Boll. Com. Glac. It. **20**.
Trevisan, L. & E. Tongiorgi. 1958. La Terra (Torino).
Troll, C. 1942. Der Büsserschnee (Nieve de los penitentes) in den Hochgebirgen der Erde. Peterm. Mitt., Erg. Heft 240.
Troll, C. 1944*a*. Diluvialgeologie und Klima. Geol. Rundschau. **34:** 307–325.
Troll, C. 1944*b*. Strukturböden, Solifluktion und Frostklimate der Erde. Geol. Rundschau. **34:** 545–694.

VAN DIJK, D. C. 1959. Soil features in relation to erosional history in the vicinity of Canberra. C.S.I.R.O. Soil Pub. 13.

VOOUS, K. H. & J. G. VAN MARLE. 1949. The distributional history of *Coracina* in the Indo-Australian archipelago. Bydr. tot de Dierk. **28:** 513–529.

WEXLER, H. 1953. Radiation balance of the Earth as a factor in climatic change. *In* Shapley.

WHITLEY, G. P. 1947. The Fluvifaunulae of Australia. W. Australian Naturalist. **1:** 49–53.

WILLETT, H. C. 1949. Long-period fluctuations of the general circulation of the atmosphere. J. Meteorol. **6:** 34–50.

WILLETT, H. C. 1949*b*. Solar variability as a factor in the fluctuations of climate during geological time. *In* Glaciers and Climate, Geogr. An. 1949.

WILLETT, H. C. 1953. Atmospheric and oceanic circulation as factors in glacial-interglacial changes of climate. *In* Shapley, 1953.

WOMERSLEY, J. S. & J. B. MCADAM. 1957. The forests and forest conditions in the territories of Papua and New Guinea (Port Moresby).

WUNDT, W. 1944. Die Ursachen der Eiszeiten. Geol. Rundschau. **34:** 713–747.

ZEUNER, F. E. 1945. The Pleistocene Period. London, England.

Part VI. Paleotemperatures and Cycle Effects

FAUNAL EVIDENCE FOR PLEISTOCENE CLIMATES

F. E. Zeuner

Institute of Archaeology, University of London, London, England

Although the paleoclimatic sequence of the Pleistocene will always be dependent chiefly on the investigation of processes of sedimentation, erosion, and denudation, fauna often provides valuable information and, in some cases, this is far superior to information obtained from geological evidence alone. Much of our knowledge of the average temperatures prevailing during the subdivisions of the Tertiary, as well as of the Pleistocene, is based on association of species in fossil faunas. It is usually assumed that these species, when identifiable with recent forms, had the same climatic requirements in the past as they have today. This is true in a large number of cases, but not in all. Elementary mistakes were made in the past as, for instance, in the middle of the last century, when elephants and lions found in the Pleistocene of Europe were both regarded as representatives of a warm climate. It required abundant evidence of the association of these species with other members of the fauna (in the case of the mammoth with reindeer and of the lion with horses, red deer, and other temperate species) to make clear that generalizations of this type are sometimes erroneous. This is especially true of groups in which evolutionary processes took place at a fast rate, such as the mammals. The opposite is likely to be the case in groups that evolve slowly at the present time, for instance, marine mollusca. Climatic deductions based on these are likely to be more reliable from the point of view of their climatic stability. On the other hand, they suffer a good deal (and even more than mammalian remains) from redeposition. One of the best examples known to me is the cliffs of early Pleistocene marine Crags which, on the coast of East Anglia, often collapse in such a way that their contained fossils are embedded in the present beach together with Recent species. That this must have happened repeatedly in the past is evident, and it is likely that all the Crag deposits contain a certain number of shells derived from older stages. In such faunas the newcomers are more significant than the survivors (Zeuner, 1937*b*). Apart from the difficulties thus caused by sedimentary processes, the correct identification of the species and subspecies is a matter of extreme importance. It is conceivable, for instance, that *Cyprina islandica*, so frequently quoted as evidence of cold conditions in the Sicilian stage of the Mediterranean, is not the same form as that which occurs now in the north Atlantic. Whilst this summary estimate of the character of faunas will always remain a useful approach to the problem of past climates, it can by no means be regarded as infallible.

For this reason, it is desirable that more research be carried out on the morphological, anatomical, and physiological functional characters of species that make it possible for them to live under certain definable climatic conditions and, indeed, often compels them to do so exclusively (Zeuner, 1936). Most groups are not sufficiently sensitive to be significant, and it is therefore necessary

to look for special cases. Two will be quoted here: one from the mammals and one from the insects.

The Relation Between Form of Skull and Manner of Life of the Rhinoceroses

The Pleistocene and Recent rhinoceroses afford a good example for the close connection between the shape of organs and mode of life (Zeuner, 1934*a*). It can be shown that the Recent species of rhinoceroses are accustomed to carry their heads in different ways in accordance with their usual manner of feeding. The average carriage of the head is exhibited by individuals when they are walking or standing at ease, although this is not the position in which the head is carried most frequently in the course of life. The skull is constructed to conform to this average carriage, as may be seen in TABLE 1.

TABLE 1

	Species			
	sondaicus	*unicornis*	*bicornis*	*simum*
Biotope	Virgin forest	Virgin forest and jungle	Forest steppe	Steppe
Food	Foliage	Foliage and grasses of the jungle	Foliage and grass	Grass
Average carriage of the head	Horizontal	Horizontal	Somewhat inclined	Strongly inclined
Angle of occiputal crest (*O*)	94°	82°	70°	64°
Angle between occiput and palate (*PO*)	45°	48.5°	67°	88°
Angle between vertical axis of Foramen magnum and palate (*Y*)	74.5°	71°	90°	110.5°

The angles employed to describe the shape of the skull are: the angle over the occipital crest (*O*), which is smaller the more the crest is extended backwards; the angle between palate and occiput (*PO*); and that between foramen magnum and palate (*Y*), both of which vary in the same direction but opposite to *O*. They indicate the extent to which the face is inserted below the cranial part of the skull.

The Recent species show clearly that forest forms have small occipital crests and jaws inserted more or less in front of the brain case (*R. sundaicus* and *R. unicornis*, the Indian rhinoceros), while in the open country forms (*R. simum*), the skull is of the "hanging type," with facial portion attached distinctly below the cranium. These differences are evident from TABLE 1, and they imply that forest rhinoceroses carry their heads more or less horizontally forward; those of forest-steppe (*R. bicornis*), somewhat inclined; and those of grasse-steppe, strongly inclined. The skull is constructed accordingly. There is sufficient evidence for this law to be regarded as generally applicable, and it enables one to argue from Recent to fossil forms. In this way it is possible to obtain information about the biotope and climate in which the extinct species was

living. Of the Pleistocene rhinoceroses of Europe, the best known is the woolly species (*Coelodonta antiquitatis*), and the average figures for skull measurements have been found to be $O = 54°$, $PO = 98°$, and $Y = 95°$. A special feature of this species is a combination of a very large PO and a much less wide Y. It finds its explanation in the fact that in this species both neck and head were strongly stretched when feeding, indicating that the food must have been very low-growing. The body of a female, complete with skin, preserved in a Pleistocene oil swamp in the Carpathians at Starunia, near Lvov, Union of Soviet Socialist Republics, bears this out as, of course, does other fauna and the flora found in association.

While *Coelodonta antiquitatis* thus indicates the steppe of the cool, periglacial belt of climate, two other Pleistocene species are typical of a more favorable climate. These are *Dicerorhinus etruscus* (chiefly Lower Pleistocene) and *Dicerorhinus kirchbergensis* (Merck's rhinoceros, chiefly Middle and Upper Pleistocene). Both these species are closely allied to each other and have indices very near to those of *Dicerorhinus bicornis*, as shown in TABLE 2.

According to these results, *etruscus* and *kirchbergensis* were inhabitants of a semiopen country bearing some bushes and trees. That these were able to grow indicates that the climate must have been warmer than in the glacial phases, and there is every reason to suppose that these two species lived in the more temperate interglacial phases.

TABLE 2

	O	PO	Y
Dicerorhinus etruscus	73°	71°	98.5°
Dicerorhinus kirchbergensis	77.5°	67°	95.5°
Dicerorhinus bicornis	70°	67°	90°

This evidence is based on purely biological and anatomical arguments but, again, independent geological evidence confirms our conclusions about the climatic conditions under which these species lived, for they are found exclusively in interglacial deposits.

It appears to be possible, therefore, to use the skull structure of the rhinoceroses as a guide to the environmental conditions in the Pleistocene.

An Insect Fauna as Evidence of Glacial Climate

Starunia became well known in 1907, when the forehalf of a woolly rhinoceros was discovered. In 1929, the Polish Academy of Sciences and Letters, Cracow, Poland, carried out new excavations (Nowak *et al.*, 1930) at the same place and discovered not only the complete female, mentioned above, but also a rich glacial flora and insect fauna. Of the insects, the grasshoppers are particularly interesting, as they have made possible a determination of the average temperatures of the summer months during a glacial phase.

The fossil orthopterous fauna of Starunia contains 14 species, all of which either belong to Recent species or have very close Recent relatives. Of these,

8 are climatically highly specialized. They all prefer grass lands and, except for one species, occur at the present day on alpine meadows above the limits of the forests. Some of them are found on meadows with high grass, whilst others are found where there are small bushes. As a whole, the fauna is surprisingly uniform, and from it we may conclude that when the woolly rhinoceros was present the countryside around Starunia consisted of meadows dotted with bushes. According to the orthopterous evidence, these bushes may have been *Pinus montana*, *Rhododendron*, *Juniperus*, and *Vaccinium*.

The fossil flora agrees very well with this inference, the following forms having been found: *Betula nana* L., *B. humilis* Schrk., *Salix reticulata* L. and other willows, *Dryas octopetala* L., *Polygonum viviparum* L., *P. lapatifolium* L., *Calluna vulgaris* var. *hirsuta* Presl., *Vaccinium uliginosum* L., *Thalictrum alpinum* L., *Thymus sudeticus* Borb., *Armeria* sp., *Phaca cf. alpina* Wulf., *Taraxacum* sp., several species of *Carex*, needles of a spruce, and numerous other forms. The following mosses have been found (Szafran, 1934; Gams, 1934): *Distichium capillaceum* (Sw.), *Tortula ruralis* (L) Ehrh., *Tortula norvegica* (Web.) Lindb., *Climacium dendroides* Web. et Mohr, *Thuidium abietinum* Br. Eur., *Polytrichum juniperinum* Willd. There is thus no doubt about the biotope of the Starunia grasshoppers.

At present all the grasshopper species occur in climatic zones not favorable to the growth of forests. Nevertheless the number of species is comparatively high, being about the same as that of favorable localities in the temperate zone of Central Europe The ecological conditions of Starunia, therefore, must have been rather propitious for Orthoptera, a group that generally requires heat for its development. As purely arctic species are not represented in the fauna but are replaced by forms needing a rather high amount of heat, the summer temperature must have reached a minimum that satisfied the latter. Thus the climatic specialization of Orthoptera enables us to obtain figures for the temperature of those months of the year during which Orthoptera were living.

For this purpose it is necessary to know the minimum temperature requirements of the species most sensitive to cold, that is, *Isophya* sp. This gives us the limit below which the mean summer temperature cannot have fallen. Equally, we need to know the highest temperature that can be borne by *Podisma frigida* Bohem., the species present most sensitive to heat. This gives us the upper limit of the mean summer temperature.

Naturally, the other species lying between these two extremes could exist under the conditions determined in this way.

During the months of the year with an average temperature below 0° C., Orthoptera cannot develop; they do not hatch from their eggs unless the thermometer rises a few degrees above zero. The hatching time of most of the German species, for example, is May or late April, when the average is about 8° to 11° C. Forms more adapted to a cold climate may start their development earlier, and in accepting an average of 3° C. for them we have certainly put the limit very low, the danger of frost being considerable for the newly hatched larvae.

In TABLES 3, 4, and 5 intervals of 3° C. are chosen to avoid a too special form and to express, at the same time, the seasonal fluctuations of temperature. A

typical locality where *Isophya* lives under cold conditions, shows the kind of climate shown in TABLE 3.

Isophya being the form most sensitive to cold, these figures indicate the minimum average temperatures that must have prevailed at Starunia. On the other hand, one of the warmest places where *Podisma frigida* is found shows the climate described in TABLE 4.

The climate of Starunia cannot have been much warmer than indicated in this table. Comparing Pontresina, Switzerland, with the Riesengebirge, we find

TABLE 3

RIESENGEBIRGE, EAST GERMANY, IN THE SUDETEN MOUNTAINS, 1415 METERS ABOVE SEA LEVEL

Above 0° C.:	May–October:	6 months
Above 3° C.:	May–September:	5 months
Above 6° C.:	June–September:	4 months
Above 9° C.:	July–August:	2 months

Highest average: July 9.9° C.

TABLE 4

PONTRESINA, SWITZERLAND, 1805 METERS ABOVE SEA LEVEL

Above 0° C.:	April–October:	7 months
Above 3° C.:	May–September:	5 months
Above 6° C.:	June–September:	4 months
Above 9° C.:	June–August:	3 months

Highest average: July 11.8° C.

TABLE 5

Above 0° C.:	April or May–October:	6–7 months
Above 3° C.:	May–September:	5 months
Above 6° C.:	June–September:	4 months
Above 9° C.:	June or July–August:	2–3 months

Highest average: July 9.9–11.8° C.

a surprisingly small overlap in the climatic ranges of the two species within which the climate of Starunia must have lain (TABLE 5).

This result is in good agreement with those of numerous independent investigations bearing on the glacial climate.

On the basis of our knowledge of the Recent relatives of the fossil species, and of the ecology of the Orthoptera in general, we have come to the conclusions briefly outlined above. The probability that the results are correct is very high in a case such as this, as the results obtained with the aid of single species check each other, and there is no species whose requirements are in contradiction to the general result. Moreover, the space of time that has elapsed since the deposits of Starunia were formed is comparatively short (roughly 100,000 years), and nearly all the fossils belong to Recent species and to genera the representatives of which are highly specialized in their mode of life. The

whole fauna exhibits a uniform character, and the conclusions based on the Orthoptera of Starunia all point in the same direction. This would be quite extraordinary if some of the species had altered their climatic requirements in the meantime without differentiating their morphological characters.

References

GAMS, H. 1934. Die Moose von Starunia als Vegetations- und Klimazeugen. Starunia, No. 2, Polska Akad. Umiej. Cracow, 6 pp.

NOWAK, J., E. PANOW, J. TOKARSKI, W. SZAFER & J. STACH. 1930. The Second Woolly Rhinoceros from Starunia, Poland. Bull. Int. Acad. Polon. Sci. Lett., Cl. math.-nat., Cracow (B), 47 pp.

SZAFRAN, B. 1934. Diluvial mosses from Starunia. Starunia, No. 1, Polska Akad. Umiej. Cracow, 17 pp.

ZEUNER, F. E. 1934*a*. Die Beziehungen zwischen Schädelform und Lebensweise bei den rezenten und fossilen Nashörnern. Ber. Naturforsch Ges. Freiburg Breisgau. **34:** 21–80.

ZEUNER, F. E. 1934*b*. Die Orthopteren aus der diluvialen Nashornschicht von Starunia (polnische Karpathen). Starunia, No. 3, Polska Akad. Umiej. Cracow, 17 pp.

ZEUNER, F. E. 1936. Paleobiology and climate of the Past.-Probl. Paleont., Publ. Lab. Paleont. Moscow Univ. **1:** 199–216.

ZEUNER, F. E. 1937*a*. The Climate of the Countries adjoining the Ice-sheet of the Pleistocene. Proc. Geol. Assoc. London. **48**(4): 379–395.

ZEUNER, F. E. 1937*b*. A comparison of the Pleistocene of East Anglia with that of Germany. Proc. Prehist. Soc. London. **3**(1): 136–157.

THE EFFECT OF INTERNAL PROCESSES AND PALEOCLIMATES

L. Egyed
Eötvös University, Budapest, Hungary

The climatic features of the past periods of geohistory are recorded by some characteristic geological formations such as those of coal, salt, bauxite, the distribution of coral reefs, and glaciations.

The changes in the climate of a given area were first explained by the changing of the climate of insolation of that area due to changes in latitude brought about by polar wandering. On the other hand, part of the climatic changes were related to local influences. However, the periodical intensity variations of the formations indicative for a given set of climatic conditions suggest that the climatic changes can be connected also with the processes of tectonism. On the basis of my studies concerning the earth's interior, I propose to review some of the factors that acted upon the climate of the entire earth in dependence on the processes of tectonism and magmatism.

In the development of paleoclimatic conditions, the processes taking place in the interior of the earth may exert direct or indirect influences. In the following, an attempt will be made to analyze briefly both of these kinds of influence.

Temperature distribution, viewed as the statistical average of geological periods of time, is necessarily a function of latitude only. Accordingly the climatic zoning taking averages over geological spans of time depends on the position of the ancient latitude circles. However, even if this general statement is accepted, there arises the question whether the amount of solar energy incident upon the surface of our globe is to be considered constant and, if not, what is the most significant factor modifying it.

This problem may be partly answered by considering the results of research on the structure and dynamism of the earth's interior. It was found that every particularity of the structure and behavior of the earth and of the solar system as a whole is based on Dirac's[1] equation stating that the gravity coefficient varies inversely as a time parameter, that is, $f = x/t$ where x is a constant and t the time parameter, its present value being, according to Gilbert,[2] 4.1×10^9 years.

Concerning the planets of the solar system it can be shown that for the average orbital radius R_n of any of the planets the relation

$$fR_n = \beta_n = \text{const.}$$

holds.[3] Considering that $f = x/t$, the earth, like the rest of the planets, was closer to the sun in the past than it is now. This means that the solar energy incident upon the unit surface of the earth, for example in the Cambrian, was about 25 per cent higher than it is now. However, the radius of the earth was also smaller at that time, so that the total incident energy—supposing always that the surface radiation of the sun did not change significantly—must have been something like the present-day one. Nevertheless, as a consequence of this state of things, the contrast between the poles and the equatorial areas was sharper. On the other hand, the climatic contrasts were mitigated by several other factors: one, by the smaller continental surface,

as proved by paleogeographical data in agreement with the earth expansion theory.[4] Moreover, the period of rotation and revolution of the earth must also have been shorter, because from the assumption of expansion it follows that the radius and consequently the momentum of inertia of the earth must have been smaller. Consequently, it is to be expected that the temperature should have been warmer, more uniform and—at least around the ancient equator—more moist on the average than at present.

Another question to be treated separately is that of the arid belts of climatic history. The aridity of a given climatic period is generally considered to be indicated by the intensity of salt deposition. However, this indicator is not as unequivocal as we could wish it to be. Undoubtedly, salt formation takes place in areas where there is scarce precipitation in a favorably hot climate. Such areas occur in the high-pressure belts situated north and south of the equator: in the tropics. The air currents directed towards the low-pressure belt of the equator carry away even the small amount of moisture present; it is in this way that the arid belts around the tropics come to exist. That this was indeed true, also in the geological past, is shown by Lotze's[5] maps of geological rock salt formation, indicating that the deposition of salts always took place along well-defined circles of the earth's surface. However, the extension of the areas of salt deposition depended to some extent on the size of the free continental surfaces. So, the salt deposition is a function of world-wide transgression and regression. This will be proved in the following.

In 1931, Szádeczky-Kardoss[6] studied the intensity changes of salt deposition in the course of the earth's history and found that, on the average, the intensity of salt precipitation had increased in the course of geological time. He considered this trend due to the secondary leaching of ancient salt deposits. If, however, Szádeczky's data, given in an arbitrary unit, are plotted against the free continental surface, an almost linear correlation is found to exist: the data are seen to be arranged independently of their temporal sequence. Therefore, we may state that the average amount of salt deposition per unit time is a function of the absolute extension of the water-free areas on the earth's surface; disregarding this fact, the development of the salt deposits is characteristic only of the arid belts situated along the ancient tropics, and not of the entire climate of any period.

The relationship presented above confirms at the same time the concept of the expansion of the earth because the paleogeographical maps are supported by this close and logical correlation with salt formation. However, the decrease of the water-covered continental surface, as shown by the paleogeographical maps, is just one of the most important proofs of the expansion of the earth's surface, that is, of the whole of its volume.

Let us now attack another problem. It is usual to relate the coldness of the climate of a given period to the intensity and extension of glaciations. According to geological observations, glaciation was intense in the Precambrian. On the other hand, glaciations can be demonstrated after each of the main phases of mountain building.

The intense Precambrian glaciation becomes intelligible only if we assume that at that time the earth's axis was almost perpendicular to the plane of

the ecliptic, because this made the contrast between the equator and the poles sharpest. The air currents outside the tropics were, in the Precambrian, highly laden with moisture. One reason for this was that extended areas were covered by water; also, the heat volume communicated by insolation was greater even at the higher altitudes. Because of the slight inclination of the

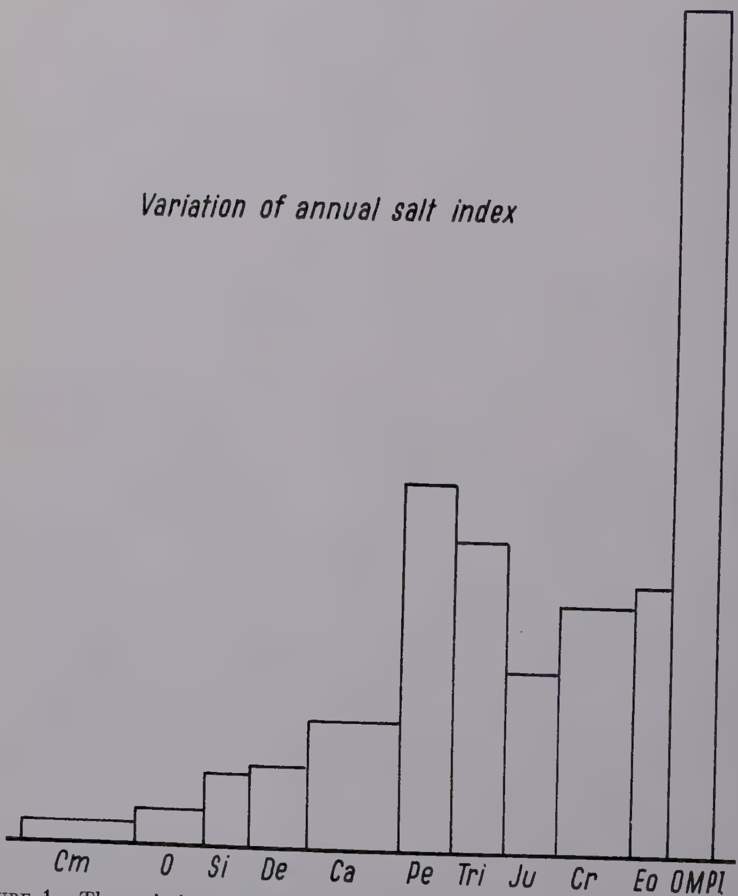


FIGURE 1. The variation of the annual salt index as given by Szádeczky in the course of the earth's history.

earth's axis, the environment of the poles had fulfilled the conditions of glaciation: that of mild winters and cool summers. The moisture-laden air masses had precipitated their moisture content in these areas, so that a gradual accumulation of snow and ice took place around the poles. This situation remained unchanged until there occurred significant changes in the inclination of the axis, or until the poles came to be situated in oceanic areas assuring a better distribution of heat. Even a somewhat greater inclination of the axis—nevertheless slighter than the present one—could have been compensated by

the fact that the year and day were shorter than they are today, as explained above. Thus the relatively long period of extensive glaciation in the Precambrian becomes understandable by the sole assumption of small axis inclination. Consequently, the Precambrian glaciation is seen to be in no connection at all with internal processes.

The problem of the glaciations succeeding the periods of mountain building is a much more difficult one to explain. The formation of high mountain chains logically results in glaciation only in the areas of the highest peaks, which could hardly have affected the faraway polar areas. The significant

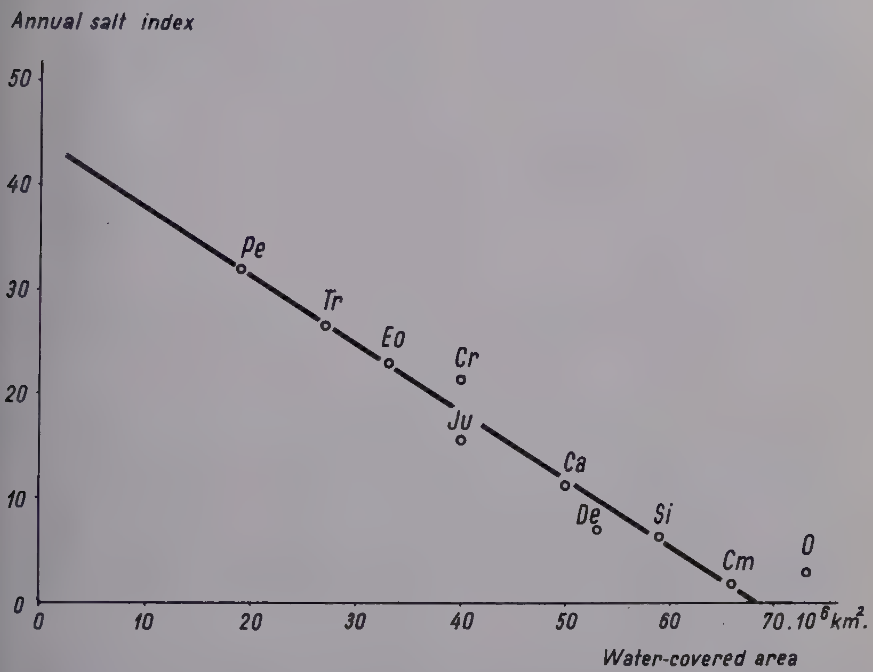


FIGURE 2. The relation of the variation of the annual salt index to the variation of the water-covered continental areas in the course of the earth's history.

effect must be looked for in the explanation of mountain building by the expansion theory. According to that theory, mountain building is connected with the formation of new oceanic areas. However, by this process, hot rock masses of a temperature close to the melting point come to exist beneath the bottom of the new ocean. These hot masses increase the heat flow to an extent estimated sufficient to warm up the water of the oceans by some degrees. Consequently, the same amount of solar energy brings about a more intense evaporation. Thus, subsequent to mountain building, an increasingly humid climate must be taken into consideration. This effect is enhanced only by the formation of shallow seas due to regression accompanying the emergence of the new mountain chains. Consequently the air masses moving towards the pole carry a greater concentration of moisture, resulting in more intense

precipitation. If this process is combined with a fluctuation of insolation similar to that proved by Milankovič, one that may be assumed for any geological period, the phases of cool summers and mild winters suffice to bring about the accumulation of snow and ice out of the available precipitation and to cause the commencement and extension of a glaciation. Consequently Milankovič's radiation pattern only plays the part of a trigger force, and the development of a glaciation is impossible without the appropriate amount of precipitation out of air masses highly laden with moisture.

There is another sedimentological climatic indicator that merits special attention, namely limestone formation. The scarcity of limestone in the Precambrian and just afterward may be due not so much to climatical influences as to the higher pressure of the ocean water due to the higher gravity coefficient and the smaller earth radius. The higher pressure had intensified the dissolution of carbonic acid in the hydrosphere; that is why limestone precipitation is the weaker the further back we proceed in the geological past. Perhaps the decrease of the pressure to a critical value permitting the escape of carbonic acid from the oceans was the factor eliciting the sudden blooming of plant life in the Carboniferous, and the precipitation of huge amounts of limestone in the Mesozoic.

In summary we may state that the climate of the earth is governed primarily by its position relative to the sun, and that the effects of the internal processes on the climates of the geological past have been mainly indirect ones.

References

1. Dirac, P. A. M. 1938. *Proc. Roy. Soc.* London. **165:** 199.
2. Gilbert, C. 1956. Dirac's cosmology and the general theory of relativity. M.N.R.A.S. **116:** 684–690.
3. Jordan, P. 1952. Schwerkraft und Weltall. Braunschweig.
4. Egyed, L. 1956. The change of the earth's dimensions determined from paleogeographical data. Geofisica pura e appl. **33:** 42–48.
5. Lotze, F. 1938. Steinsalz und Kalisalze. Berlin, Germany.
6. Szádeczky-Kardoss, E. 1931. A sóképzödés intenzitásváltozásai / Variations in the intensity of salt formation. Földtani Közlöny. **60:** 34–57.

CORRELATION OF BEACH TERRACES WITH CLIMATIC CYCLES OF PLUVIAL LAKE LAHONTAN, NEVADA

Richard Shutler, Jr.
Nevada State Museum, Carson City, Nevada

My colleagues and I are attempting to develop methods to compliment stratigraphy and radiocarbon dating in correlating Lake Lahontan beach terraces with the climatic fluctuations responsible for their formation. The levels of Pyramid Lake in historic times are discussed to provide additional evidence that climate, that is, precipitation and evaporation, is responsible for the rises and falls of the Lahontan lakes even in recent times. Emission spectrography and thermoluminescence have been applied to tufa with promising results for climatic interpretations. The presence of pollen in tufa and the lake bottom and terrace sediments will permit application of palynology to derive climatic information from them. Sedimentation, geomorphology, and archaeology are also useful tools in this climatic study.

Radiocarbon dating has established a chronology for the major rises and falls of Lake Lahontan in prehistoric times (Broecker and Orr, 1958). It seems worthwhile to explore the historical data available on the precipitation in the Sierra Nevada Mountains, and the information on the elevations of Pyramid Lake since its discovery in 1844 by John C. Fremont. This approach is designed to tell us if our attempt to correlate the ancient Lahontan beach terraces (lake levels) with climatic cycles is based on the same variables that control the fluctuations of Pyramid Lake in recent times.

This is best done by a study of the water-supply and precipitation conditions in the Truckee River Basin in relation to the levels of Pyramid Lake during the past 100 years.

Taylor (1902) made an attempt to correlate the river discharge with volume of water evaporated from the lake, but did not complete the study. Antevs (1938) developed curves of the volumes of water in Pyramid and Winnemucca lakes, but apparently did not account for the considerable quantities of water used within the Truckee River Basin and diverted to points outside it after 1907. In 1941 Hardman and Venstrom brought together all the available information on the changes that have taken place in the levels of the two lakes since about 1840. Their estimates are based on changes that would have occurred under undisturbed natural conditions. This was done by computing the amount of water that entered the Truckee River Basin that never reached Pyramid Lake. For our purpose here, I shall refer only to the data pertaining to Pyramid Lake as the variables that determined the levels of Winnemucca Lake are more complicated, and are not pertinent to the present problem.

Pyramid Lake lies about 40 miles northeast of Reno, Nev. It is approximately 30 miles long and from 8 to 12 miles wide.

Correlation of the Truckee River runoff with the levels of Pyramid Lake is possible because there is a general relationship between the precipitation in the San Francisco-Sacramento area and the precipitation east of the Sierra Ne-

vadas. Weather records have been available for the stations of San Francisco and Sacramento since 1849–1850. This record has been used as a general basis of moisture conditions in the Truckee River Basin since 1850.

The Truckee River derives the major portion of its flow from the high elevations of the Sierra Nevada Range. For this reason the precipitation of the mountains may be used in deriving an index of the runoff of the river. However, caution must be exercised in using this index, as there appear to be instances in which the precipitation in the Sierras does not indicate the runoff accurately. This is because in certain years the precipitation may be relatively heavier on the lower portion of the watershed than on the high mountains and, in some years, the reverse is true. To obtain a more accurate measurement of the precipitation on the entire watershed, Hardman and Venstrom used the information from collecting stations in the Truckee River Basin as well as that from the Sierra stations to compile their index. Our discussion of the levels of Pyramid Lake is based on information derived by this method.

A summary of the information gathered from historical records and other sources by Hardman and Venstrom on the history of Pyramid Lake is presented in the following paragraphs:

"Pyramid Lake was slightly below the overflow-level into Lake Winnemucca in 1844, and continued low until about 1860. A period of rather abundant moisture, accompanied by heavy floods in the Truckee River in 1862, 1868, and 1869, resulted in a high level which continued to 1871 . . . it appears clear that the level for 1871 was not over 3876 feet. Extreme high levels of 3879 feet are assumed to have been reached in 1862, and 1868 or 1869. The lake fell about 15 feet from an elevation of 3876 feet in 1871 to an elevation of 3861 feet in 1889, then is reported by Jones (1925) to have risen 17 feet, to 3878 feet, in the year 1890. Pyramid Lake remained near the level of Winnemucca Slough until about 1917, since which time it has fallen rather steadily. 'These variations in lake level were due to the varying amounts of precipitation in the Truckee River Basin.' The drop in the level of 'Pyramid Lake' was accelerated by the diversion of water to the Carson River Basin following the completion of the Truckee Canal in 1907. In the drought-period of the late 1920s and 1930s diversions for irrigation within and outside the basin took the great bulk of the total flows. The combined effort of declining river-flow and of increased irrigation-demands caused Pyramid Lake to drop about 50 feet from its normal level by the spring of 1938 to an elevation of 3816 feet," which was the lowest point reached since its discovery in 1844 until the present. The elevation of Pyramid Lake today is approximately 3800 feet.

In spite of the present use of Truckee River water for projects other than Pyramid Lake, it is clear that the precipitation on the high and low elevations of the Truckee River Basin, was, until disturbed by man, one of two principal factors responsible for the levels of Pyramid Lake, and by inference, ancient Lake Lahontan. The other important factor to be considered is evaporation. It has been estimated that Pyramid Lake loses about 52 and 54 inches annually by evaporation. On the basis of historical records alone, then, it can be shown there is a correlation between climate (precipitation and evaporation) and the levels of one of the two remnants of Lake Lahontan, Pyramid Lake.

At its maximum elevation of 4383 feet above sea level about 11,700 years ago, Lake Lahontan occupied 20 interconnected basins in western Nevada and northeastern California. Its greatest length north and south was 240 miles, while its east-west limit was 172 miles. The surface area of the high lake was approximately 8000 square miles. The lake was never a solid body of water. It surrounded land masses which in turn contained contemporary lakes. Pyramid and Walker lakes are the present day remnants of this once large body of water.

The hydrographic basin covered about 40,775 sq. mi. The Susan, Truckee, Carson, and Walker rivers from the west and the Humboldt River from the east flow into this area of internal drainage.

Research on Lake Lahontan problems began in the late 1800s; such workers as Russell (1885), Jones (1925), Antevs (1925, 1938, 1952), Hutchinson (1937), Hubbs (1948), Morrison (n.d.), and Broecker and Orr (1958), have been the principal investigators. The major lake rises and falls have been broadly defined and dated. This latter work by Broecker and Orr (1958), have also confirmed the conclusions of other workers that climatic variations were the cause of the lake fluctuations. The lake levels depending on the precipitation rate for each basin, rate of evaporation, and amount of melt water received from mountain glaciers of the Sierra Nevadas and upper reaches of the Humboldt river.

The traditional explanation for the cause of these pluvial lakes is that during the glacial periods the ice sheets in Canada and in the northern United States forced the pressure and precipitation belts south of their present positions. The Pacific anticyclone did not develop to block summerly cyclones, so there was probably summer precipitation. The cold and rainy season was longer. The Pacific Ocean and North American continent experienced pressure changes causing different cyclonic storms to enter the continent at different latitudes, therefore the precipitation was dispersed over the entire region south of the glaciers during the winter months. During this period the Great Basin pluvial lakes must have experienced their time of greatest precipitation. Reduced evaporation as a consequence of lowered temperature would be a very important contributing factor of the greater amount of moisture.

The high level lakes dated by Broecker and Orr (1958) at 11,700 and 9500 to 10,000 years ago, formed terraces over the entire Lahontan area that were due to the same climatic factors, as during this time most all of the basins were filled. After the lake level fell to the point where there became established about 20 separate basins during the last 9000 years, it would seem probable that each basin followed a separate history. Local changes in a basin caused by a river or other microclimatic factors may affect the one basin only. We now have evidence to support this contention.

Our efforts at the Nevada State Museum are directed towards establishing a correlation of Lake Lahontan terraces with climatic fluctuations as represented on the lower elevations of each basin. We are concerned with the lower elevation terraces because many of them have evidence of human occupancy on their surface, and other evidence suggests that a uniform series of events are responsible for the high level terraces. This being the case, it is the

factors that were responsible for the formation of the low level terraces which need clarification. At the present time no evidence of man has been found on shore lines above the 4200 foot level. Our broad program includes an attempt to interpret the human prehistory of the Lahontan area, reconstruct the past environment, and correlate the shore lines with the climatic fluctuations that were so important to their formation. These problems are all interrelated. Since there is evidence that the basins have had separate histories since the lake level fell below the 4200 to 3950 foot level, it would seem logical to work out the climatic history of a number of the basins before attempting area-wide correlations of the low-level terraces.

Our first task was to find an area within a basin that exhibited as many terraces as possible to use as a pilot study, and then to determine if some of the techniques named above are applicable to tufa, the calcium carbonate deposit found on these terraces and the lake bottom sediments now represented as playas. We already know that radiocarbon dating may be applied with valid results to tufa (Broecker and Orr, 1958).

To ensure the validity of such results, we felt that by first applying these methods to the high elevation tufas and by obtaining similar results on tufas from different basins, since these were supposedly deposited under similar conditions, we could then expect reasonably valid results from the low elevation tufas, the assumption being that diverse results would indicate diverse conditions of deposition.

An area known as Boot Hill was located on the north side of Humboldt Sink, about 5 miles southwest of Lovelock, Perishing county, Nev., by M.M. Wheat, who is doing the preliminary geological field work.

This area of about 11½ sq. mi. embraces a cuspate bar that has recorded all but the lowest 75 feet of the shore lines of the Carson-Humboldt basin, or the upper 430 feet of Lake Lahontan. The highest Lake Lahontan terraces are exhibited here, down to the 3980-foot terrace.

The shark's-tooth shape of the cuspate bar was maintained through all the fluctuations of Lake Lahontan, so that on the northeast or depositional side of the bar many successive offshore bars and longshore troughs have been developed. These have in turn buried or have been buried by other bars as the lake elevation rose and fell.

On the erosional side to the southwest, the ends of the bars have been truncated, leaving a cross section of older bars exposed. The material thus obtained by wave action has been carried by later and lower wave action along the base of the spit to form a new bar. On these bars, which became terraces after the recession of the water level, was deposited an impure calcium carbonate known as tufa. Tufa, the principal subject of our study to date is of 5 main types: thinolite, a crystalline chemical precipitate not found above 4030 feet elevation; dentritic, deposited by algae growing in tufted mats; lithoid, deposited either by chemical precipitation or smooth algal mats is the tufa found at the highest elevation of any; cellular, deposited by tufted algal mats under strong wave conditions stopping abruptly at 4335 feet altitude; and coralline, similar to the cellular except that it is deposited in deeper, quieter water.

We determined the absolute elevation of about 18 terraces in relation to the tufa on them and collected samples of these tufas. Our purpose in doing this was to test the tufas for climatically sensitive trace elements and for the presence of pollen. We have been successful in both instances. The samples are sufficient for eventual radiocarbon dating.

It should be pointed out here that Russell (1885) conceived of the possibility that the chemical composition of the tufas might give a clue to the conditions under which they were deposited. He collected samples from different elevations and from different basins, and these were subjected to chemical analyses. This work, performed by O.D. Allen, did not prove fruitful. "Chemical composition of the tufas gives no hint as to differences in the conditions under which they were deposited" (Russell, 1885, p. 203).

Bruno E. Sabels of the Institute of Geophysics, University of California at Los Angeles, Calif., has developed a method of paleoclimatic investigation by quantitative trace-element analysis. He has found manganese to be the most reliable climatic indicator in fossil feces and soils in cave strata. The manganese is easily soluble in the acidic soil solutions of wet, cool climates, but is almost insoluble under alkaline, dry, and warm desert conditions. Therefore plants and their products, feces and ashes, reflect the climate during their growth period by their manganese content. Sabels found that desert plants and feces contain 100 to 600 ppm manganese, while plants and feces from glacial environments in present day desert areas contain 1000 to 1700 ppm manganese, with transition stages in between. In analyzing cave strata, pollen and radiocarbon data are used as supporting tools, as demonstrated in our study of Rampart Cave in the Grand Canyon (Martin, Sabels, and Shutler, in press).

Tufas from Boot Hill and from Pyramid and Winnemucca lakes dated by Broecker, and provided by Orr, were studied by this trace-element method. Three elements, iron, manganese, and magnesium, were found to have potential climatic interpretation value. This work was done in the spectrography laboratory of the Department of Geology, University of California at Los Angeles. The thermoluminescence technique was also applied to the tufas with apparent success. Sabels is trying to develop this method for the dating of pottery with George Kennedy, under a grant from the National Science Foundation, Washington, D.C.

Thermoluminescence is a system for dating materials that have at some time been heated to 400° C. or higher. By reheating the material to 400° C. and measuring the proton emission, the "glow effect" or thermoluminescence, an estimate of the length of time that elasped since the material was last heated to that temperature can be made. Although tufa has not been heated to 400° C., the application of this method to tufa in relation to climatic change is that a large "glow" accompanies tufa deposited during a humid period, and a small "glow" from tufa deposited during an arid period.

In testing tufa for the presence of pollen I ran a sample of the tufa presently being deposited in Pyramid Lake and samples of tufa from three different elevations at Boot Hill. The results of the tests established the presence of pollen in all four. Identification of the pollen has not been completed, but

I did note several different types on the slides. We are aware of the problems of wind and current distribution yet to be solved before results can be properly evaluated and are working on them.

A gravel pit has been opened in the lake bottom sediments at the foot of Boot Hill. I took samples of these sediments from a 30-foot profile at this location. The palynology laboratory of the Geochronology Laboratories at the University of Arizona, Tucson, Ariz., determined that these sediments contain pollen. No pollen count has been made, but pine pollen was numerous on some of the slides. This indicates that the coring of similar playas will produce pollen-bearing samples from these lake-bottom sediments. We hope that this procedure will permit us to obtain samples going back into the early history of Lake Lahontan, and to establish possible correlation with the corresponding tufas found on the terraces.

There are no archaeological relics on the Boot Hill terraces, nor are there dry caves or rockshelters in this part of the Humboldt Sink, but I propose to describe the important part archaeology can play in the climatic-environmental studies of Lake Lahontan. Hundreds of campsites found on terraces contain a variety of stone tools, and numerous dry caves and rockshelters occur throughout the region. The variety of perishable artifacts found in these dry sites in the Lahontan basin and the types of tools associated with the terrace campsites give clues to the adjustment to the environment and use of natural resources by man who, we know, has been in the area for at least 8000 years. Analysis of the material culture, together with the floral and faunal remains, and pollen and trace element studies of the deposits of these sites provide evidence of the past climatic conditions in the area, which can be radiocarbon dated. Two examples of trace element studies on archaeological deposits in the Lake Winnemucca area will illustrate the point. Sabels analyzed soil samples from a dry cave at an altitude of approximately 4250 feet. I had taken the samples at 6-inch levels down 138 inches in depth during excavation of the cave. There was 11½ feet of culture in this site. Using the ppm of manganese in each sample Sables interprets the occupation and past climate of the area as having been occupied by man through the last part of an extensive cold period with high Lake Lahontan levels, which most probably was the late Wisconsin glacial age. We shall be able to check this interpretation with radiocarbon dates and pollen analysis. At a rockshelter on the same hill, based on soil samples from a 3-foot profile, he interprets the site as having been occupied only during the intermediate or late period of life in these sites, based on intermediate to low manganese values.

Extinct fauna also plays a part in temporal and climatic studies. On the bottom of this same rockshelter we have found the lower jaw and part of the upper jaw of the extinct *Euceratherium*, a shrub ox related to the musk ox. This is the first find of this animal in Nevada. This genus is known from the mid- and late Pleistocene in some caves in northern California and other areas of North America. A find of the same genus at Burnett Cave in New Mexico was assumed to be associated with a radiocarbon date of 7432 years B.P. Other extinct fauna such as horse and camel have been found in the archaeological sites of the Winnemucca Lake area, indicating that man was contempo-

raneous with these now extinct fauna, and that he has been in the area for a considerable period of time. Many older, pre man fossils are to be found in the strata of the Lahontan basin, so paleontology also plays its part in the research.

I have discussed or mentioned the tools we have to work with in attempting to correlate the shore lines of Lake Lahontan with the climatic fluctuations that caused them. My remaining remarks will consider briefly what we have accomplished at Boot Hill and how we propose to proceed with the problem.

Three Boot Hill tufa samples and two radiocarbon dated tufa samples L 364 AA and L 289 G (9500 $\pm$ 200; 9700 $\pm$ 200) from the same range of elevation, between 4320 and 4380 feet were found to have similar manganese contents of 17 to 35 ppm. Significantly, these samples had similar, large glow areas. The good agreement in elevation, composition (Mn, Fe) and glow area is important as the samples from Boot Hill and those radiocarbon-dated came from locations about 100 miles apart. At high water levels, fresh water conditions apparently lead to similar conditions in widely separated parts of the Lahontan Basin.

Comparison of tufas from lower elevations at Boot Hill and from similar elevations at Pyramid Lake has shown that with low water levels the similarity, or identity, of samples from different parts of the basin but of the same elevation is nonexistent. While Mn and Fe in the Pyramid Lake region seems to decrease slightly with falling lake level and increasing dryness, they increase significantly in the Boot Hill area. From these facts it seems apparent that in arid times individual lakes with their separate chemical variations, 8 to 16 ppm Mn, and different levels affected by rate of the influx of water, rate of evaporation, and other variables will have followed separate histories.

Cool, wet climate tufa show little impurities and, as a consequence, large thermoluminescence glow. Warm, dry climate tufa, formed in brackish water at low lake levels, show abundant impurities and small glow areas. Accordingly tufa can be assigned to the climate present during its formation by using trace elements and thermoluminescence. It appears that there is a correlation between glow size and manganese abundance in relation to climate. While the two methods have given essentially the same results with this study, we are aware that there are other pitfalls to resolve. Although the correlation between glow size and manganese abundance in relation to the climatic fluctuations in the Lahontan Basins appears to be established, we plan to check these results with radiocarbon dating and pollen analysis. However, if our hypothesis is correct, we shall not expect the low level terraces from basin to basin to correlate in time and climatic conditions.

The general conclusions of the climatic studies on Lake Lahontan to date are:

(1) Radiocarbon dating has established a chronology for the major rises and falls of Lake Lahontan.

(2) Historical records of the past 100 years have shown that it is climatic change (precipitation in the Sierra Nevada's and the Truckee River Basin, and evaporation from the lake surface), that is responsible for the varying levels of Pyramid Lake in recent times.

(3) Lack of evidence of human occupancy on Lahontan beach terraces above the 4200-foot level is probably due to the fact that man was not in the area until the climate changed in the high mountains and the lake level fell below a certain elevation.

(4) Archaeology can play an important part in the climatic-environmental studies of Lake Lahontan by conducting analyses of the material culture and of floral and faunal remains in sites, and by carrying out pollen and trace-element studies of soil samples from the deposits.

(5) Paleontology, through the study of fossils and extinct fauna recovered from archaeological sites, provides evidence for climatic interpretations.

(6) Trace element and thermoluminescence studies are two new methods of deriving climatic information from tufa.

(7) Trace elements (principally manganese) and thermoluminescence indicate that similar conditions are responsible for the formation of the high Lahontan lake levels, while the low-level lakes in the various basins have followed separate histories.

(8) The presence of pollen in tufa and the lake bottom sediments will provide additional evidence of the climate present in the Lahontan Basin for the past several thousand years.

(9) Evidence accumulated by the techniques described or mentioned here leave little doubt that climatic cycles were responsible for the changing levels of Lake Lahontan and its remnants, Pyramid and Walker lakes.

Building on previous research and adding the methods discussed here, we feel our pilot study at Boot Hill will eventually provide a sound, productive approach to our goal of correlating the low level beach terraces with the climatic fluctuations of pluvial Lake Lahontan, Nev., by bringing together these converging lines of evidence.

References

ANTEVS, E. 1925. On the Pleistocene history of the Great Basin. Carnegie Inst. Wash. Publ. **352:** 53–114. 1938. Rainfall and tree growth in the Great Basin. Carnegie Inst. Wash. **469**. 1952. Cenozoic climates of the Great Basin. Geologische Rundschau. **40:** 94–108. 1955. Geologic-climatic dating in the west. Am. Antiquity. **20**(4): Pt. 1, 317–335.

BROECKER, W. S. & P. C. ORR. 1958. Radiocarbon chronology of Lake Lahontan and Lake Bonneville. Bull. Geol. Soc. Am. **69:** 1009–1032.

HARDMAN, G. & C. VENSTROM. 1941. A 100-year record of Truckee River runoff estimated from changes in levels and volumes of Pyramid and Winnemucca lakes. Trans. Am. Geophys. Union.

HUBBS, C. L. & R. R. MILLER. 1948. The Great Basin. With emphasis on glacial and postglacial times. Bull. Univ. Utah. **38**(20): 17–166.

HUTCHINSON, G. E. 1937. A contribution to the limnology of arid regions. Primarily founded on observations made in the Lahontan Basin. Trans. Conn. Acad. Arts Sci. **33:** 47–132.

JONES, J. C. 1925. Geologic history of Lake Lahontan. Carnegie Inst. Wash. **352:** 3–50.

RUSSELL, I. C. 1885. Geological history of Lake Lahontan. A Quaternary lake of northwestern Nevada. U. S. Geol. Survey. **11**.

TAYLOR, L. H. 1902. Water storage in the Truckee Basin, California-Nevada. Dept. Intern. Geol. Survey W-S. **68**.

CENOZOIC CLIMATIC CHANGES AS INDICATED BY THE STRATIGRAPHY AND CHRONOLOGY OF DEEP-SEA CORES OF GLOBIGERINA-OOZE FACIES*

Cesare Emiliani

The Marine Laboratory, University of Miami, Coral Gables, Fla.

Temperatures in the atmosphere, at the earth's surface, and in the oceans have changed during geologic time following wide spectra of duration, amplitude, and areal extent. Without prejudice as to symmetry or periodicity, temperature changes may be termed cyclical because conditions similar to the original ones are re-established sooner or later.

Local fluctuations over periods of years or even decades may result from random fluctuations in the atmospheric or oceanic circulation. Regional and world-wide temperature fluctuations of similar duration are often ascribed to solar causes, especially sunspot cycles and associated variations in the ultraviolet, X-ray, and corpuscular components of solar emission. Local and regional fluctuations over periods of millions of years are related to the dynamics of local and regional land physiography. As an example, temperature at the 500 mb. level above the Colorado Plateau has certainly changed as a result of the uplift of the plateau during the Cenozoic, and a return to the original conditions may be expected only after the plateau itself will have been dissected and eroded away. World-wide temperature fluctuations lasting tens of thousands of years (the glacial-interglacial cycles of the Pleistocene) have been ascribed to a complex mechanism involving both astronomical and terrestrial factors. Finally, world-wide temperature cycles lasting tens of millions of years have been ascribed either to changes in solar emission or to changes of the earth's albedo associated with orogenesis and marine transgression.

The evidence for climatic change during geologic time may be summarized as follows:

(1) A number of major epochs of glaciation have occurred, spaced in time 100 to 300 million years (TABLE 1).

(2) The last two glacial epochs, in the Late Paleozoic and Late Cenozoic, occurred in times of strong marine regression, increased continentality, and marked orogenesis.

(3) In nonglacial times, as between the Late Paleozoic and the Late Cenozoic, the latitudinal temperature gradient was smaller than during glacial epochs, and temperature fluctuations of very long duration and of more than regional significance occurred. Thus oxygen isotopic analysis of belemnites of Late Cretaceous age from Alaska and Siberia have shown that surface temperature of the sea water at high latitudes was about 14° C. at that time (Epstein, 1959). In addition, oxygen isotopic analysis of belemnites and other fossils from Europe and North America have shown the occurrence, during the Late Cretaceous, of a broad temperature cycle having an amplitude of about 5° C., and a duration of about 30 million years (FIGURE 1; Lowenstam and Epstein, 1954; Bowen, 1961). Systematic temperature fluctuations of shorter

* Contribution No. 315 from the Marine Laboratory, University of Miami.

duration (tens of thousands of years) have not been detected in sediments deposited during nonglacial ages (Emiliani, 1956) (FIGURES 2 and 3).

(4) A broad temperature decrease occurred during the Tertiary, amounting to about 8° C. for the middle latitudes. The evidence is both paleontological

TABLE 1

SUMMARY OF THE EARTH'S GLACIAL EPOCHS, THEIR AGES, AND THE AFFECTED AREAS*

Glacial epoch	Time (10^6 years)	Main glaciated areas
Pleistocene	0.6 to 0	North America; central and northern Europe; northern Siberia; Alps; Himalaya; Andes; Patagonia; New Zealand; Tasmania; Antarctica
Late Paleozoic	250	South America; Central and South Africa; Madagascar; India; southern Australia
Late Pre-Cambrian	600	Spitzbergen; Greenland; Scandinavia; Australia; Yang-tze, China (?)
Transvaal-Nama-Katanga	700	South West Africa; Angola; Congo; Rhodesia; Transvaal; Simla, India (?); Broken Hill, New South Wales (?)
Huronian	800	Cobalt, Ontario; Witwatersrand, Transvaal
Bothnian	1000	Finland; Western Australia (?)
Damara	1200 (?)	Chuos District, South West Africa
(No name)	1500 (?)	Medicine Bow Mountains, Wyo.

* Partly after Holmes, 1937.

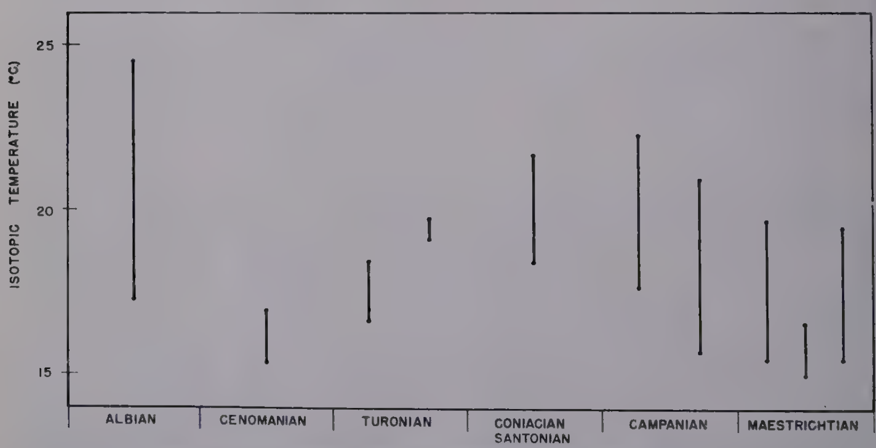


FIGURE 1. Temperature ranges during various epochs of the Cretaceous, obtained by average oxygen isotopic analysis of belemnites from western Europe (data from Lowenstam and Epstein, 1954).

and paleobotanical, and covers both Europe and North America (cf. Emiliani, 1954, and references therein).

(5) Bottom temperature of the equatorial Pacific and, by inference, surface temperatures in the polar seas, decreased about 8° C. during the Tertiary, as indicated by oxygen isotopic analysis of calcareous benthonic Foraminifera (Emiliani, 1954). If these data are complemented with inferences from the

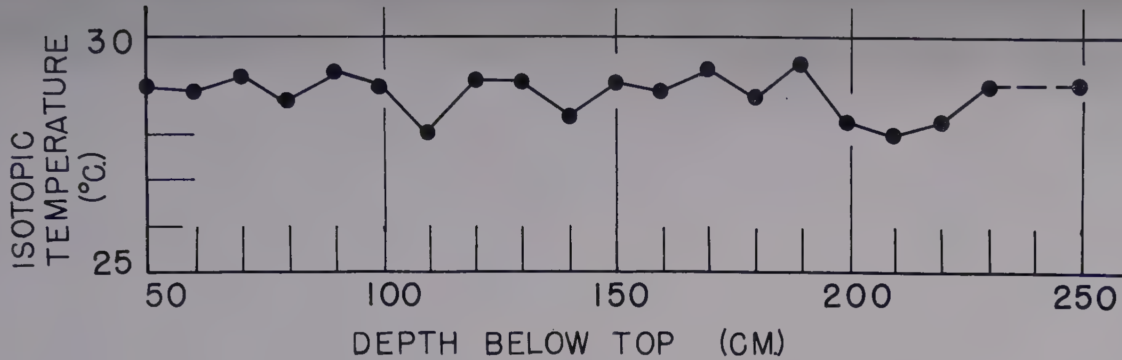


FIGURE 2. Surface temperature of the subtropical Atlantic during a portion (about 100,000 years) of the Middle Oligocene, obtained by oxygen isotopic analysis of shells of the pelagic foraminifer *Globigerinoides sacculifera*. Reproduced by permission of the *Journal of Geology* (Emiliani, 1956).

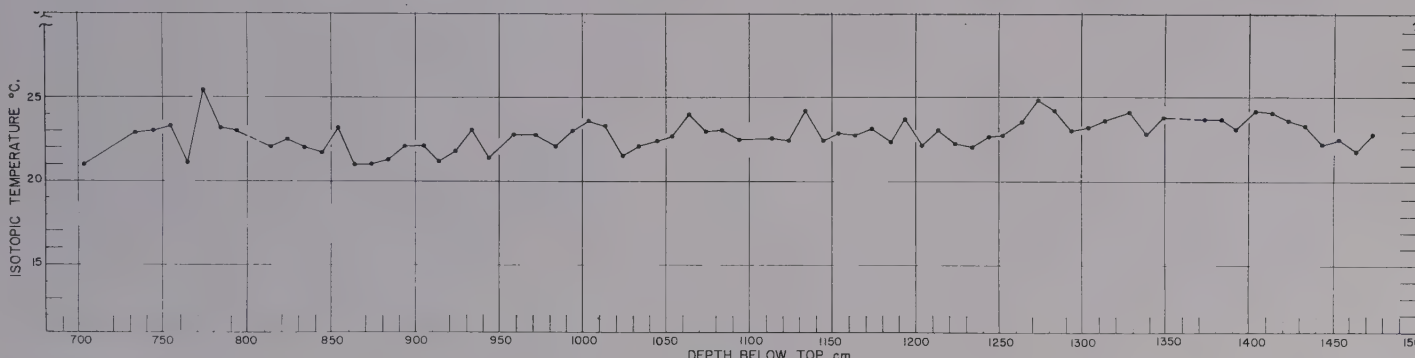


FIGURE 3. Surface temperature of the equatorial Atlantic during a portion (about 300,000 years) of the Lower-Middle Miocene, obtained by oxygen isotopic analysis of shells of the pelagic foraminifer *Globigerinoides sacculifera*. Reproduced by permission of the *Journal of Geology* (Emiliani, 1956).

Alaskan and Siberian belemnites mentioned above, a temperature decrease of about 12° C. appears to have occurred both in surface waters at high latitudes and in bottom water in all open oceanic basins during the past 75 million years (FIGURE 4). Although the available data are few and the chronological control poor, the decrease does not seem to have been linear with time but to have become faster during the Cenozoic. A correction for the possibly greater concentration of O^{18} in sea water in the Late Cretaceous, partly balanced by the absence of icecaps, may raise the amount of the decrease by 1 or 2° C. (cf. Emiliani, 1955, p. 541).

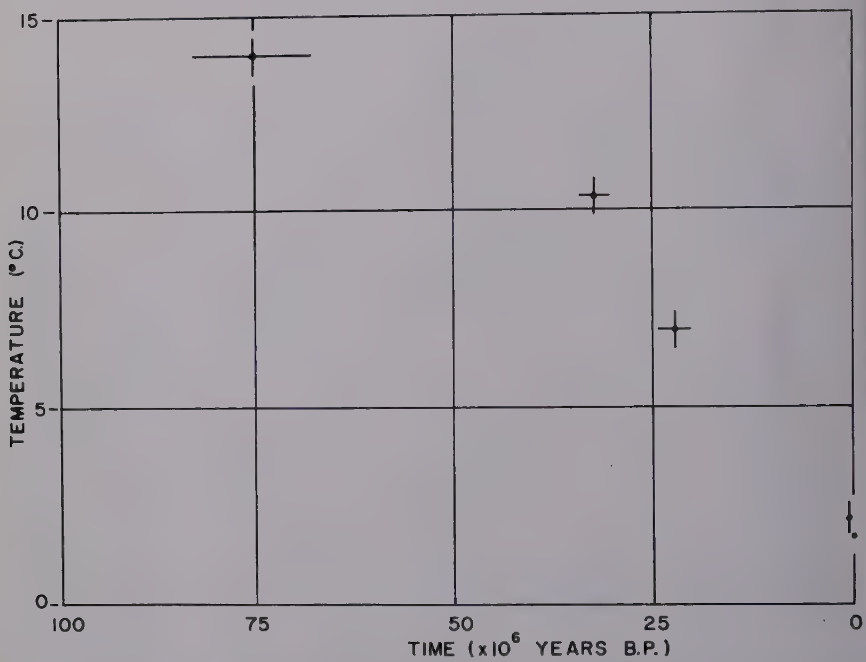


FIGURE 4. Temperature decrease of the surface water at high latitudes and of the bottom water in open oceanic basins during the past 75 million years.

(6) Bottom temperature of the eastern equatorial Pacific was already similar to the present in Late Pliocene time (FIGURE 4), suggesting that glaciation of Antarctica, and perhaps also of Greenland, was already complete, and ice extended to the sea shores.

(7) Sometime within the past million years temperature began to fluctuate, the fluctuations being measured in tens of thousands of years rather than tens of millions of years.

(8) Rapid temperature fluctuations were occurring already in the Pliocene, but became greater in the Early Pleistocene and later. In the Mediterranean, the secular temperature minima decreased from 21° C. in the Pliocene to 16° C. in the Early Pleistocene and to 12° C. in the Later Pleistocene. These values have been obtained by oxygen isotopic analysis of pelagic Foraminifera from

the Plio-Pleistocene section at Le Castella near Crotone, southern Italy (Emiliani *et al.*, 1961) and from a deep-sea core from the eastern Mediterranean (Emiliani, 1955*a*). They are believed to represent mean summer temperatures of the surface sea water.

(9) Temperature fluctuations persisted during the Pleistocene, as shown by alternating tills and soils in glaciated areas, alternating loess and soils in periglacial areas, alternating salt and clay deposits in certain lakes, alternating warmer and colder pollen assemblages in many lacustrine deposits, alternating warmer and colder faunas of pelagic Foraminifera in deep-sea cores of Globigerina-ooze facies, and alternating higher and lower O^{18}/O^{16} ratios in the shell assemblages of these Foraminifera.

(10) Within the Pleistocene, temperature cycles of more than regional significance ranged in duration from some tens of thousands of years to a few thousands of years and probably much less.

Geologists working on continental glacial and periglacial deposits have recognized four or five major glacial ages within the Pleistocene epoch, separated by major interglacials. The first major glaciation is believed by many to have coincided with the beginning of the Pleistocene, fixed at the time when certain species of northern marine invertebrates appeared in the continuous Plio-Pleistocene sections of Italy (International Geological Congress, 1950). This coincidence, however, is still only hypothetical.

Inferences as to the number and chronology even of only the major temperature fluctuations of the Pleistocene epoch are difficult to make on the basis of glacial and periglacial deposits. These deposits, in fact, are always discontinuous and fragmentary and cannot be correlated with each other by fossil content because Pleistocene time has been too short to allow significant speciation of the lower organisms that generally leave more abundant fossils. Some sections of continental deposits, seemingly representing the whole Pleistocene, are known from nonglacial areas. These are the lacustrine sections described by Clisby *et al.* (1957) and by Eardley and Gvosdetsky (1960). Unfortunately, the mineralogical and palynological record of these sections is difficult to interpret unequivocally in terms of Pleistocene stratigraphy.

Continuous sections of deposits, some of them covering the whole Pleistocene and more, have been recovered from the deep-sea floor by means of the piston corer invented by Kullenberg (1947). Sections of Globigerina-ooze sediment, containing large amounts of pelagic foraminiferal shells, are particularly useful for the study of Pleistocene temperatures. These temperatures can be estimated from the relative abundances of different species of pelagic Foraminifera having different temperature tolerances (Philippi, 1910), or can be measured by the oxygen isotopic method devised by Urey (1947) and developed by Urey, Epstein, and their co-workers (Epstein *et al.*, 1951, 1953). Oxygen isotopic analysis of pelagic Foraminifera yields the temperature of the sea water at or near the surface, at the time and place where the foraminiferal shells grew. Calculation of the temperature values from the isotopic ratios in fossil shells requires, however, an estimate of the oxygen isotopic composition of the sea water. Although such estimates can be made with reasonable confidence for open oceanic waters of the past, the temperature values obtained cannot be considered as unequivocally determined.

In spite of this difficulty, isotopic analysis of deep-sea cores has yielded a picture of changing temperature that promises to be of much help in clarifying the sequence and chronology of Pleistocene events.

Paleotemperature analysis of Swedish deep-sea core 58 from the eastern equatorial Pacific, containing a complete section of Pleistocene sediments and some of the Pliocene sediments below (Arrhenius, 1952), shows a temperature decrease of about 3° C. from the Late Pliocene into the Early Pleistocene, followed by temperature oscillations of about 3° C. of amplitude during the Pleistocene (FIGURE 5; Emiliani, 1955). The latter do not provide a clear temperature record for the Pleistocene because of the local pattern of vertical circulation of the surface water, because only species of deeper habitats were available for isotopic analysis (cf. Emiliani, 1954*a*) and because the amplitude of the temperature variation was small (Emiliani, 1955). Numerous oscillations in the calcium carbonate percentage in core 58 and other cores from the same region have been noticed (Arrhenius, 1952). These oscillations are best interpreted in terms of climatic changes, although the exact relationship is not clear. The Plio-Pleistocene boundary can be variously placed at 600 to 800 cm. below the top of core 58 on the basis of the paleotemperature record alone, although a correlation with the Plio-Pleistocene sections of Italy, on which the definition of Plio-Pleistocene boundary rests, has not yet been made. Fifteen to 20 carbonate cycles occur in the Pleistocene sections of cores 58 and 62, suggesting the occurrence of an equivalent number of climatic oscillations. Only some of these apparently resulted in major glaciations.

A far clearer picture of climatic change during the Pleistocene is provided by oxygen isotopic analysis of deep-sea cores from the Atlantic Ocean and adjacent seas (FIGURE 6; Emiliani, 1955, 1955*a*, 1958; Rosholt *et al.*, 1961). Reasons for the greater clarity are the greater amplitude of the temperature variations, probably in response to more direct heat exchange with the northern icecaps, and the availability of superficial species of pelagic Foraminifera for isotopic analysis. The temperature graphs of the various cores (FIGURES 6 and 7 and FIGURE 1 in Emiliani, 1955*a*) show that:

(1) Temperature oscillated between maxima and minima.

(2) The maximum temperature range is 6 to 7° C. in the equatorial Atlantic and about 12° C. in the Mediterranean, if a correction is made to account for the greater concentration of O^{18} in ocean water during glacial ages (see Emiliani, 1955).

(3) Most maxima and minima reach equivalent values, but some (FIGURE 6, stages 3, 8, 12) do not.

(4) The background noise is apparently about 2° C. (FIGURES 6 and 7), introduced by (a) imperfect mixing of the sediment by bottom animals; (b) sampling statistics related to the fact that not more than a few hundred shells are used for each isotopic analysis; (c) occurrence within each sample for isotopic analysis of shells of somewhat different sizes, although almost always larger than 250 μ (cf. Emiliani, 1955, p. 548), possibly representing slightly different depth or seasonal growth (cf., however, Emiliani, 1954*a*); (d) the analytical error (±0.5° C.).

(5) The amplitude of the major temperature cycles is several times greater

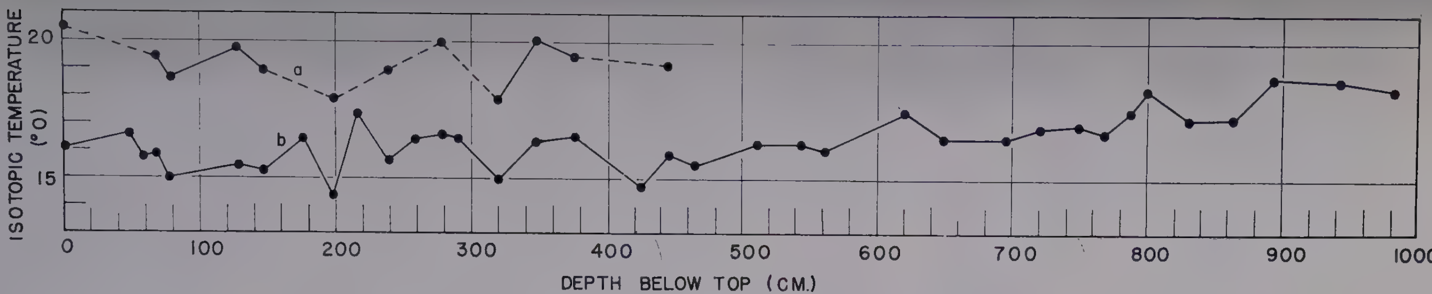


FIGURE 5. Core 58, eastern equatorial Pacific: isotopic temperatures obtained from the pelagic Foraminifera *Pulleniatina obliquiloculata* (*a*) and *Globorotalia tumida* (*b*). Reproduced by permission of the *Journal of Geology* (Emiliani, 1955).

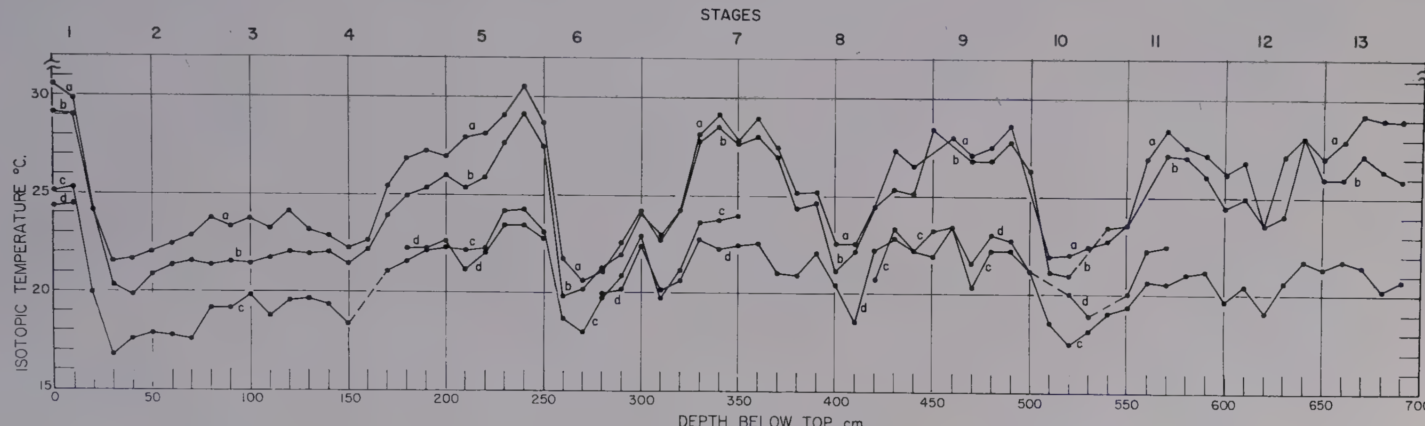


FIGURE 6. Core A179-4, central Caribbean: isotopic temperatures obtained from (*a*) *Globigerinoides rubra;* (*b*) *Globigerinoides sacculifera;* (*c*) *Globigerina dubia;* and (*d*) *Globorotalia menardii*. Reproduced by permission of the *Journal of Geology* (Emiliani, 1955).

than noise, so that these cycles stand out clearly and would be apparent also if a single core had been analyzed.

(6) The amplitudes of the minor temperature oscillations compare with noise. Therefore detection of these oscillations requires filtering, which is done simply by intercorrelating two or more cores. As an example, the minor temperature maximum at 600 cm. below top in core 280 and at 350 cm. below top in core 234 (FIGURE 7) is believed to represent a true temperature episode

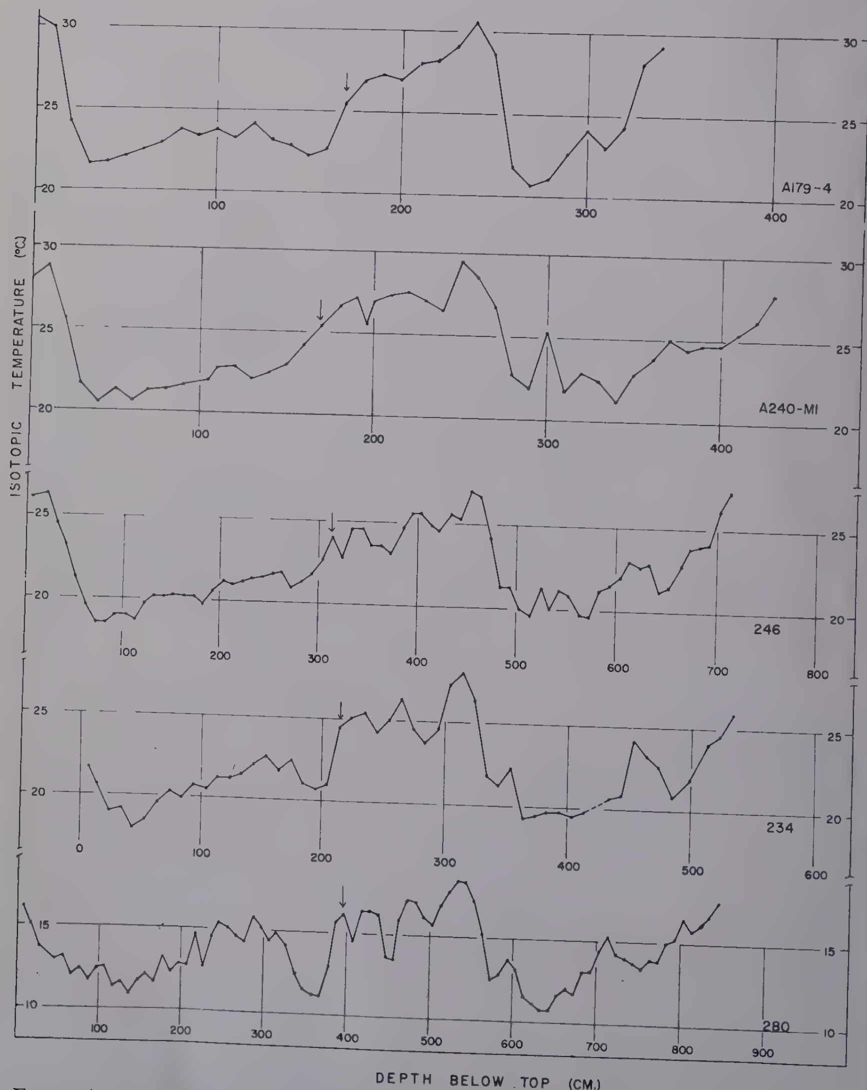


FIGURE 7. Isotopic temperature curves of various cores. Core numbers to the right. Core 280, north Atlantic; cores 234 and 246, equatorial Atlantic; cores A240-Ml and A179-4, Caribbean. Arrows indicate last occurrence of the subspecies *Globorotalia menardii flexuosa* Reproduced by permission of the *Journal of Geology* (Rosholt *et al.*, 1961).

because it appears in equivalent stratigraphic position in two different cores (which are about 4000 km. apart).

The chronology of the temperature oscillations rests on C^{14} measurements by Rubin and Suess (1955, 1956) and on Pa^{231}/Th^{230} measurements by Rosholt (Rosholt *et al.*, 1961). C^{14} analysis showed that the last low-temperature stage lasted from about 27,000 to 12,000 years ago, and clearly established a correlation with the Main Würm-Main Wisconsin glaciation of Europe and North America. Pa^{231}/Th^{230} analysis confirmed the C^{14} chronology and, in addition, dated earlier temperature maxima and minima at 55 to 65 thousand years, 95,000 years, 106,000 years, and about 135,000, 150,000, and 175,000 years (FIGURE 8).

The temperature curves of the various deep-sea cores have been combined into a single, generalized temperature curve (FIGURE 9). Warm stages are identified by odd integers and cold stages by even integers, all increasing with age. The curve is time-calibrated by C^{14} and Pa^{231}/Th^{230} dating back to 175,000 years B.P. (Rubin and Suess, 1955, 1956; Rosholt *et al.*, 1961), and by extrapolating beyond. C^{14} and Pa^{231}/Th^{230} analysis has dated stage 2 at 12 to 30,000 years, stage 3 at 30 to 50,000 years, stage 4 at 50 to 65,000 years, stage 5 at 65 to 100,000 years, stage 6 at 100 to 125,000 years, stage 7 at 125 to 145,000 years, stage 8 at 145 to 165,000 years, and stage 9 at 165 to 180,000 years. Comparison with the C^{14} chronology of continental Pleistocene shows that stage 2 represents the last major glacial age, the Main Würm or Main Wisconsin; stage 3 the Laufen; stage 4 the Early Würm or Early Wisconsin; and stage 5, by inference, the last interglacial (cf. Flint and Brandtner, 1961).

Correlation between earlier temperature stages of the deep-sea cores and glacial and interglacial stages on the continents is still open. Thus Zeuner (1959) correlated stage 6 with the Early Würm and Ericson and Wollin (1956) and Ericson *et al.* (1961) correlated stages 7 to 13 with the last interglacial. Note, however, that the temperature minima of the earlier cold stages are as low as the temperature minima of stages 4 and 2, which correspond to the Early Würm and the Main Würm on C^{14} evidence. Therefore, unless geologists working on continental Pleistocene deposits have made major errors, that is, unless they have failed to recognize major cold periods within interglacial ages, the earlier cold stages of the deep-sea cores should correspond in full to earlier glaciations. If so, the Riss glaciation would be dated at 100 to 125,000 years, the Mindel at 180 to 210,000 years, and the Günz, by extrapolation, at 265 to 290,000 years. It is important to notice here that, while temperature maxima and minima are probably synchronous everywhere in the world (cf. Emiliani and Geiss, 1959), the boundaries between high- and low-temperature stages may be placed at different times in different regions. Glaciation, in fact, as recognized by glacier movements and other features, may begin earlier and end later at higher latitudes than further away from the poles. *This observation has an important bearing on detailed Pleistocene correlations.* For instance, the two minor temperature oscillations following the temperature maximum of the last interglacial could perhaps also be assigned to the Early Würm glaciation.

Preliminary K/A dating of Pleistocene volcanics associated with glacial deposits has yielded ages of about 400,000 years for the Mindel and about 1 million years for an early glaciation in California (Evernden *et al.*, 1957; Evernden,

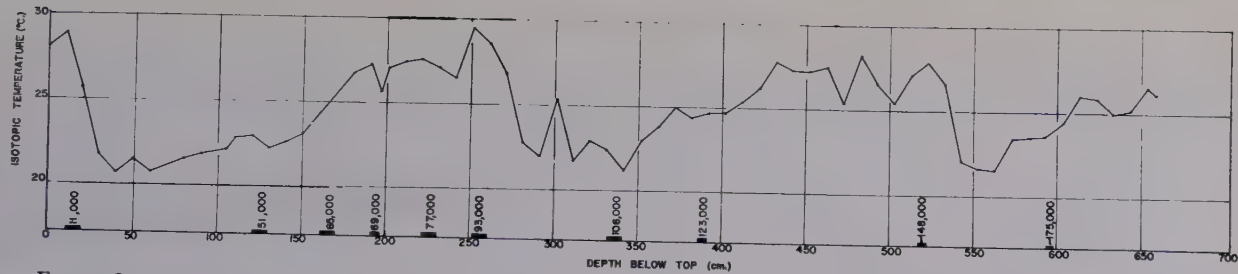


FIGURE 8. Core A240-Ml, central Caribbean: isotopic temperatures and absolute dates by the Pa^{231}/Th^{230} method. Reproduced by permission of the *Journal of Geology* (Rosholt *et al.*,1961).

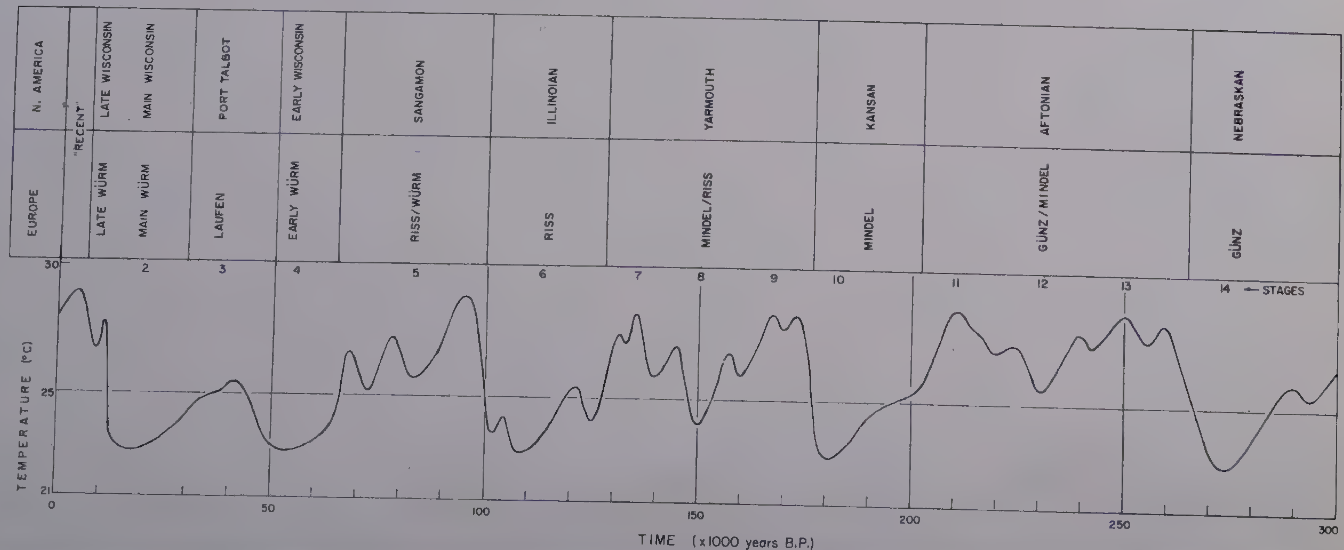


FIGURE 9. Generalized temperature curve and absolute time scale. The time scale beyond 150,000

1959). These dates are in marked disagreement with the dates mentioned above. Possible explanations for the discrepancy include errors in the correlations shown in FIGURE 9 for the time preceding the last interglacial; and doubts with respect to the validity of the K/A dates, K/A dating of Pleistocene volcanics being still in the experimental phase.

Within each major glacial and interglacial age, temperature oscillations of shorter duration occurred. Oscillations shorter than a few thousand years, such as those accompanying the substages of the Main Wisconsin, are generally not resolved by deep-sea cores. Benthonic animals, in fact, continuously rework the sediment to a depth of a few to several centimeters, corresponding to a time of a few thousand years. Therefore deep-sea cores are useful for studying temperature oscillations that lasted several thousand years or more (Emiliani, 1957). These include not only the major glacial-interglacial cycles, but also important oscillations within these cycles (FIGURES 6 and 7). Similarly, the sequence of loess and soils in certain critical continental areas contain evidence not only of the major glacial-interglacial cycles, but also of smaller cycles within the major ones. If the composite loess and soil sequence in Austria described by Brandtner (1954, 1956) is compared with the temperature record obtained from the deep-sea cores (FIGURE 10), a close relationship is apparent (Emiliani, in press).

Broecker *et al.* (1958, 1960) have marshalled evidence suggesting that an abrupt temperature rise may have occurred within a few hundred years about 11,000 years ago. The deep-sea cores, generally, do not show such an abruptness, probably a result of sediment reworking by bottom animals. Disappearance of sea ice in the northern North Atlantic and elsewhere, causing a decrease of albedo, may explain the rapidity of the temperature rise (Emiliani and Geiss, 1959). Sea ice, in fact, is thin and can be melted away very rapidly.

The evidence discussed above has some bearing on theories of glaciation. There are, of course, many such theories, some of them attempting to explain the occurrence of major glacial epochs, and others specializing in the explanation of major temperature oscillations within the Pleistocene. Variations in portions of solar emission have been proposed to explain temperature oscillations of shorter durations, that is, up to a few hundred years, and variations in Pettersson's tidal force have been proposed to explain temperature oscillations lasting a few thousand years (Karlstrom, 1957).

The occurrence of major glacial epochs has been ascribed to variations in solar emission inherent in the mechanism of nuclear reactions by which solar energy is produced (Öpik, 1958), or to geotectonic effects tending to increase the albedo of the earth. The latter explanation is favored by the apparent coincidence of both the Late Paleozoic and Late Cenozoic glaciations with times of marked marine regression, increased continentality, and intense orogenesis, while warmer climates prevailed at times of marine ingression (for example, Jura-Cretaceous).

Theories attempting to explain temperature variations within glacial epochs, in particular the alternation of glacial and interglacial ages during the Pleistocene, may be divided into two groups: those that are theoretically possible and those that are not. The latter include theories in which glaciation is controlled by deglaciation without any time delay. Thus in an early attempt to explain

deglaciation, I suggested freezing of the northern North Atlantic and the Norwegian Sea as a cause (Emiliani, 1957*a*). Similarly, Ewing and Donn (1956, 1958; Ewing, 1960) suggested freezing of the Arctic. In both cases, glaciation

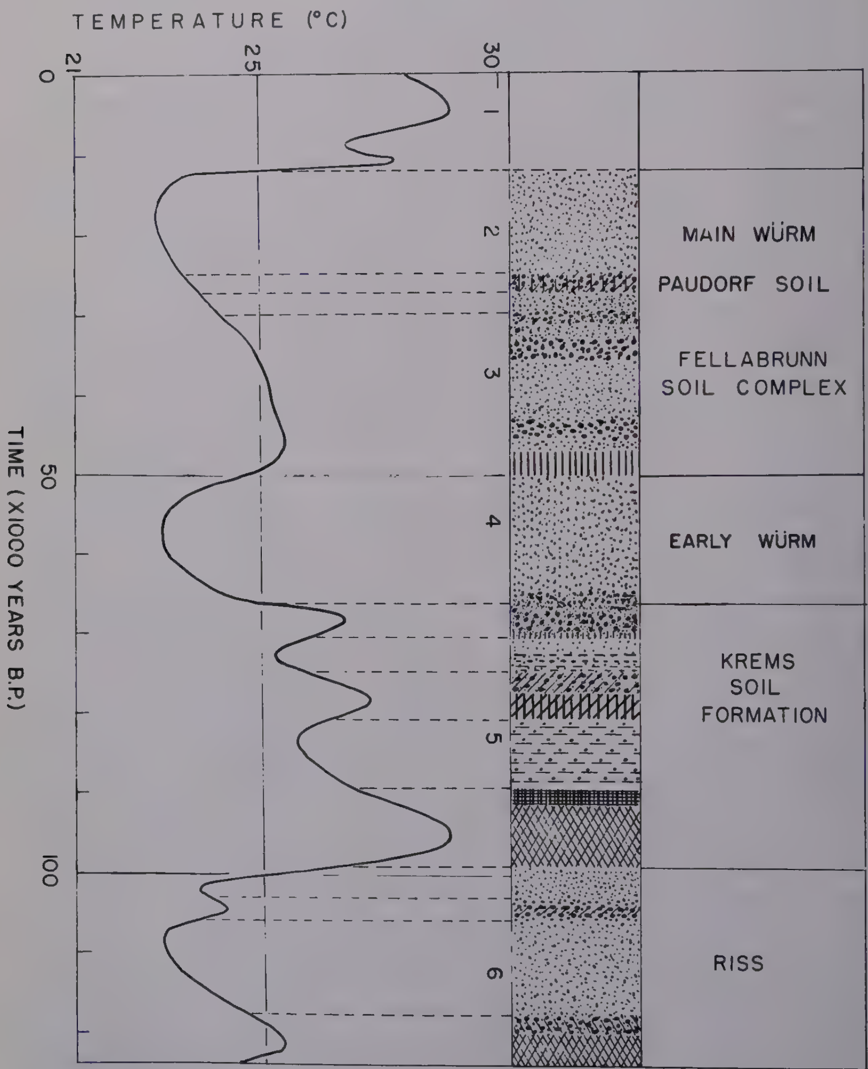


FIGURE 10. Correlation between isotopic temperatures from deep-sea cores and loess and soil profiles from Austria and Moravia (from Emiliani, in press).

and deglaciation are controlled by immediate heat exchange between continental ice and the ocean (through the atmosphere in one case, through the ocean in the other), and the very first act to start glaciation is also the very first act to stop it. In both cases, therefore, it is clearly impossible to start and maintain a sustained oscillation (Emiliani and Geiss, 1959; Livingstone, 1959)

and the only possible outcome is an equilibrium situation with the coexistence of a certain amount of continental and sea ice. Any apparently supporting evidence for such theories must be regarded as purely coincidental.

Theories that are theoretically possible can be divided into two groups: those that are contradicted or made questionable by the available evidence, and those that are not. Most recent theories fall in the first group. Thus Simpson's theory (Simpson, 1934, 1940, 1957) that surface temperatures at low latitudes during glacial ages were alternately lower and higher than during intervening interglacials is incompatible with the observed or inferred synchronism between glacial ages and the temperature minima of the deep-sea cores. The group of theories based on volcanism (Flint, 1957, p. 504–505, and references therein) and on submarine tectonism along the Greenland-Iceland-Scotland axis (Saks *et al.*, 1955) are contradicted by the apparent periodicity of glaciation: there are no reasons at present to believe that endogenous phenomena might have been periodic. Theories based on variations of the CO_2 content of the atmosphere (Plass, 1956; also Flint, 1957, p. 504) are made improbable by the rapid exchange of CO_2 between atmosphere and ocean, including the deep-sea (Craig, 1957; Revelle and Suess, 1957). Theories based on variations of solar emission are contradicted by Öpik's conclusion, based on elaborate, quantitative analysis (Öpik, 1953), that the reduction of solar emission required to produce the observed ice sheets of the glacial ages would have reduced the equatorial temperature to 8° C.: such drastic temperature decrease would have killed all equatorial and tropical fauna and flora.

According to a model of glaciation developed by Emiliani and Geiss (1959) and largely based on ideas by earlier authors (esp. Milankovitch, 1938; Köppen and Wegener, 1924; Flint, 1947), glaciation was made possible by Tertiary cooling caused by albedo increase associated with the Alpine orogenesis and marine regression; the successive Pleistocene glaciations were started by quasi-periodic reductions of summer insolation in northern latitudes (the so-called Milankovitch mechanism), known to have occurred at intervals of about 40,000 years; further development of glaciation was largely a self-sustaining process; and complete deglaciation outside of Antarctica and Greenland resulted from surface freezing of the northern North Atlantic and time-delay effects introduced by plastic flow of glaciers, heat absorption by ice melting, and crustal warping.

This model of glaciation is the only one which, thus far, has not been questioned on theoretical grounds or disproved by any known fact. On the other hand, it cannot be considered as proved either, although C^{14} and Pa^{231}/Th^{230} dating appear to provide some strong support. Absolute dating of more Pleistocene material antedating the last interglacial and significantly related to climatic changes will provide a final test for this and other theories of glaciation. Meanwhile, if this theory is correct, a marked cooling or a full glaciation should be well under way within a few thousand years, and world temperature should reach another minimum on or about 20,000 A.D.

Summary

Temperatures in the atmosphere, at the earth's surface, and in the ocean have changed during geologic time following wide spectra of duration, ampli-

tude, and areal extent. Causes for temperature changes range from variations of solar emission to changes in land physiography, marine movements, the astronomical motions of the earth, random fluctuations in atmospheric and oceanic circulation, etc. Without prejudice as to symmetry or periodicity, temperature changes may be termed cyclical, because conditions similar to the original ones are re-established sooner or later.

The major, world-wide climatic events in the history of the earth involve (1) major epochs of glaciations, occurring at intervals of 100 to 300 million years; (2) temperature cycles lasting a few to many millions of years during nonglacial times; and (3) repeated temperature cycles lasting tens of thousands of years during the Pleistocene. Geological evidence for these temperature changes rests on a variety of observations and measurements, involving many fields, including those of paleontology, paleobotany, geomorphology, geochemistry, and pedology. Oxygen isotopic analysis of Belemnites and other fossils revealed the occurrence of a temperature cycle during the Late Cretaceous having an amplitude of about 5° C. and a duration of about 30 million years. Similar analyses on Cretaceous Belemnites from Alaska and Siberia and on Tertiary benthonic Foraminifera from the equatorial Pacific showed that surface temperatures in high latitudes and bottom temperatures in all open oceanic basins decreased about 12° C. during the past 75 million years.

The study of long deep-sea cores of Globigerina-ooze facies from the Atlantic Ocean and adjacent seas using classical micropaleontological methods, determination of the O^{18}/O^{16} ratio in shell assemblages of pelagic Foraminifera, and absolute dating by C^{14} and the ratio Pa^{231}/Th^{230} provides a good insight into the broader climatic trends of the Pleistocene epoch. Thus it has been possible to clarify the repeated occurrence of temperature oscillations; to date the more recent temperature minima at 20,000, 60,000, 110,000 and 150,000 years ago; to date earlier temperature minima, by extrapolation, at about 180,000, 230,000 and 275,000 years ago; and to correlate the major temperature oscillations with the glacial and interglacial stages of the continental stratigraphy. Correlations for the time preceding the last interglacial are only tentative but, if correct, an age of about 300,000 years would be assigned to the Günz glaciation. Correlations between the base of the Pleistocene Series, defined on the Plio-Pleistocene sections of southern Italy, and deep-sea cores, has not yet been established. On the basis of paleotemperature analysis alone, the Plio-Pleistocene boundary may be placed between 600 and 800 cm. below the top of core 58 from the eastern equatorial Pacific, corresponding to an estimated age of 600 to 800,000 years.

Increasing knowledge on the sequence and chronology of Pleistocene events has disproved or made improbable most theories of glaciation proposed to date. Still unquestioned is a model of glaciation (Emiliani and Geiss, 1959) involving increase of earth albedo during the Tertiary, time control of Pleistocene glaciations by the so-called Milankovitch mechanism, and time-delay effects introduced by plastic ice flow, heat absorption by ice melting, and crustal warping.

References

Arrhenius, G. 1952. Sediment cores from the east Pacific. Swedish Deep-Sea Exped. 1947–1948, Repts., **5** (1): 227.

BOWEN, R. 1961. Paleotemperature analysis of Mesozoic Belemnoidea from Germany and Poland. J. Geol. **69:** 75–83.

BRANDTNER, F. 1954. Jungpleistozäner Löss und fossile Böden in Niederösterreich. Eiszeitalter und Gegenwart. **4/5:** 49–82.

BRANDTNER, F. 1956. Lösstratigraphie und paläolitsche Kulturabfolge in Niederösterreich. und in den angrenzenden Gebieten. Eiszeitalter und Gegenwart. **7:** 127–175.

BROECKER, W. S., M. EWING & B. C. HEEZEN. 1960. Evidence of an abrupt change in climate close to 11,000 years ago. Am. J. Sci. **258:** 429–443.

BROECKER, W. S., K. K. TUREKIAN & B. C. HEEZEN. 1958. The relation of deep sedimentation rates to variations in climate. Am. J. Sci. **256:** 503–517.

CLISBY, K. H., F. FOREMAN & P. B. SEARS. 1957. Pleistocene climatic changes in New Mexico, USA. Geobotanisches Inst. Rübel, Zürich, Veröff. **34:** 21–26.

CRAIG, H. 1957. The natural distribution of radiocarbon and the exchange time of carbon dioxide between atmosphere and sea. Tellus. **9:** 1–17.

EARDLEY, A. J. & V. GVOSDETSKY. 1960. Analysis of Pleistocene core from Great Salt Lake, Utah. Geol. Soc. Am. Bull. **71:** 1323–1344.

EMILIANI, C. 1954. Temperatures of Pacific bottom waters and polar superficial waters during the Tertiary. Science. **119:** 853–855.

EMILIANI, C. 1954*a*. Depth habitats of some species of pelagic Foraminifera as indicated by oxygen isotope ratios. Am. J. Sci. **252:** 149–158.

EMILIANI, C. 1955. Pleistocene temperatures. J. Geol. **63:** 538–578.

EMILIANI, C. 1955*a*. Pleistocene temperature variations in the Mediterranean. Quaternaria. **2:** 87–98.

EMILIANI, C. 1956. Oligocene and Miocene temperatures of the equatorial and subtropical Atlantic Ocean. J. Geol. **64:** 281–288.

EMILIANI, C. 1957. Temperature and age analysis of deep-sea cores. Science. **125:** 383–387.

EMILIANI, C. 1957*a*. Glaciations and their causes. Committee on Research in Water Resources, University of California, Conference on Recent Research in Climatology, Scripps Institution of Oceanography. March 25–26, 1957 Proc. : 36–42.

EMILIANI, C. 1958. Paleotemperature analysis of core 280 and Pleistocene correlations. J. Geol. **66:** 201–214.

EMILIANI, C. Stratigraphie sous-marine et continentale du Pléistocène. Vth Con. INQUA, Madrid-Barcelona 1957, Actes. In press.

EMILIANI, C. & J. GEISS. 1959. On glaciations and their causes. Geol. Rundschau. **46** (1957): 576–601.

EMILIANI, C., T. MAYEDA & R. SELLI. 1961. Paleotemperature analysis of the Plio-Pleistocene section at Le Castella, Calabria, southern Italy. Geol. Soc. Am. Bull. **72:** 679–688.

EPSTEIN, S. 1959. The variations of O^{18}/O^{16} ratio in nature and some geologic implications. *In* Researches in Geochemistry. P. H. Abelson, Ed. Wiley. : 217–240.

EPSTEIN, S., R. BUCHSBAUM, H. LOWENSTAM & H. C. UREY. 1951. Carbonate-water isotopic temperature scale. Geol. Soc. Am. Bull. **62:** 417–425.

EPSTEIN, S., R. BUCHSBAUM, H. LOWENSTAM & H. C. UREY. 1953. Revised carbonate-water isotopic temperature scale. Geol. Soc. Am. Bull. **64:** 1315–1325.

ERICSON, D. B. & G. WOLLIN. 1956. Micropaleontological and isotopic determinations of Pleistocene climates. Micropaleontology. **2:** 257–270.

ERICSON, D. B., M. EWING, G. WOLLIN & B. C. HEEZEN. 1961. Atlantic deep-sea sediment cores. Geol. Soc. Am. Bull. **72:** 193–286.

EVERNDEN, J. F. 1959. (Discussion.) Geol. Soc. London, Proc., Sess. 1958–1959. : 17–18.

EVERNDEN, J. F., G. H. CURTIS & R. KISTLER. 1957. Potassium-argon dating of Pleistocene volcanics. Quaternaria. **4:** 13–17.

EWING, M. 1960. The ice ages theory. Alberta Soc. Petrol. Geol. J. **8:** 191–201.

EWING, M. & W. L. DONN. 1956. A theory of ice ages. Science. **123:** 1061–1066.

EWING, M. & W. L. DONN. 1958. A theory of ice ages II. Science. **127:** 1159–1162.

FLINT, R. F. 1947. Glacial Geology and the Pleistocene Epoch. Wiley. New York, N.Y.

FLINT, R. F. 1957. Glacial and Pleistocene Geology. Wiley. New York, N.Y.

FLINT, R. F. & F. BRANDTNER. 1961. Climatic changes since the last interglacial. Am. J. Sci. **259:** 321–328.

HOLMES, A. 1937. The Age of the Earth. Nelson. London, England.

KARLSTROM, T. N. V. 1957. Tentative correlation of Alaskan glacial sequences, 1956. Science. **125:** 73–74.

KÖPPEN, W. & A. WEGENER. 1924. Die Klimate der geologischen Vorzeit. Berlin. p. 256.

KULLENBERG, B. 1947. The piston core sampler. Svenska Hydr.-Biol. Komm., Skr., Tredje Ser., Hydr., Bd. 1, H. 2, 46 p.

INTERNATIONAL GEOLOGICAL CONGRESS, 18th Session, Great Britain, 1948 (1950). Recom-

mendations of commission appointed to advise on the definition of Pliocene-Pleistocene boundary. Rept. Part 9: 6.
LIVINGSTONE, D. A. 1959. Theory of ice ages. Science. **129:** 463–465.
LOWENSTAM, H. A. & S. EPSTEIN. 1954. Paleotemperatures of the post-Aptian Cretaceous as determined by the oxygen isotope method. J. Geol. **62:** 207–248.
MILANKOVITCH, M. 1938. Astronomische Mittel zur Erforschung der erdgeschichtlichen Klimate. Handb. Geophys. **9:** 593–698.
ÖPIK, E. J. 1953. A climatological and astronomical interpretation of the ice ages and of the past variations of terrestrial climate. Armagh Obs. Contr., No. 9, 79 p.
ÖPIK, E. J. 1958. Solar variability and palaeoclimatic changes. Irish Astron. J. **5:** 97–109.
PHILIPPI, E. 1910. Die Grundproben der deutschen Südpolar-Expedition, 1901–1093. II. Geographie und Geologie, Bd. **6:** 411–616.
PLASS, G. N. 1956. The carbon dioxide theory of climatic change. Tellus. **8:** 140–154.
REVELLE, R. & H. E. SUESS. 1957. Carbon dioxide exchange between atmosphere and ocean, and the question of an increase of atmospheric CO_2 during the past decades. Tellus. **9:** 18–27.
RUBIN, M. & H. E. SUESS. 1955. U. S. Geological Survey radiocarbon dates II. Science. **121:** 481–488.
RUBIN, M. & H. E. SUESS. 1956. U. S. Geological Survey radiocarbon dates III. Science. **123:** 442–448.
ROSHOLT, J. N., C. EMILIANI, J. GEISS, F. F. KOCZY & P. J. WANGERSKY. 1961. Absolute dating of deep-sea cores by the Pa^{231}/Th^{230} method. J. Geol. **69:** 162–185.
SAKS, V. N., N. A. BELOV & N. N. LAPINA. 1955. Our present concepts of the geology of the central Arctic. Priroda. **7:** 13–22.
SIMPSON, J. C. 1934. World climate during the Quaternary period. Roy. Meteorol. Soc. Quart. J. **60:** 425–478.
SIMPSON, J. C. 1940. Possible causes of change in climate and their limitations. Linn. Soc. London Proc. **152:** 190–219.
SIMPSON, J. C. 1957. Further studies in world climate. Roy. Meteorol. Soc. Quart. J. **83:** 459–485.
UREY, H. C. 1947. The thermodynamic properties of isotopic substances. J. Chem. Soc. **1947:** 562–581.
ZEUNER, F. E. 1959. The Pleistocene Period. Hutchinson. London, England.

PLEISTOCENE CLIMATIC RECORD IN SOME DEEP-SEA SEDIMENT CORES*

David B. Ericson
Lamont Geological Observatory, Columbia University, Palisades, N. Y.

During the past 15 years about 2000 sediment cores have been raised in the North Atlantic and adjacent seas. From these we now know that conditions of sediment accumulation vary greatly from place to place on the ocean floor and from time to time, and that uninterrupted particle-by-particle sediment accumulation over long periods of time is exceptional. In the great majority of cores in the Lamont Geological Observatory collection at Palisades, N.Y., there is unmistakable evidence of disturbance of the climatic record.

Disturbance or interruption of the climatic record may be due either to catastrophic addition of sediment layers by turbidity current deposition or to removal of some part of the record by slumping or turbidity current erosion.

As a rule intercalated layers deposited by turbidity currents are not misleading. Usually they differ strikingly from sediment of slow continuous accumulation. By sampling only those layers that have the typical features of slow accumulation it is sometimes possible to obtain a fairly good climatic record in areas where turbidity currents have been of frequent occurrence. The principal difficulty is that in such areas only short records can be obtained because of the greatly augmented rate of sediment accumulation due to the instantaneously deposited turbidity current layers.

On the other hand, turbidity current layers have climatic significance in themselves. In both the submarine delta of the Hudson River submarine canyon and in the Sigsbee deep of the Gulf of Mexico the postglacial layer of sediment is entirely or largely composed of burrow-mottled foraminiferal lutite, typical of slow accumulation. However, below this layer almost all cores enter and bottom in a nearly continuous series of graded layers deposited by turbidity currents.

In the Sigsbee deep, even the longest cores fall short of reaching the zone of low-latitude foraminifera laid down before the onset of the last ice age. Nevertheless several do pass through a zone of burrow-mottled foraminiferal lutite that, judging from its thickness, must represent a time interval of several thousand years. Although the foraminifera in this layer do not in themselves indicate an amelioration of climate, the absence of turbidity current deposition during this time does strongly suggest either higher sea level, less precipitation, or both: in other words, a climatic shift toward interstadial conditions. According to a radiocarbon age determination by Wallace Broecker, this zone was laid down 20,000 years ago (Ewing *et al.*, 1958).

Loss of some part of the record because of slumping or submarine erosion poses a more difficult problem. Immediately after removal of a layer of sediment by slumping, slow accumulation at the site recommences and, soon after that, mud feeders recolonize the area. These, by churning the sediment,

* The work reported in this paper was supported in part by Grants G-763, G-4174, and G-6540 from the National Science Foundation, Washington, D. C.; and Contribution No. 499 from the Lamont Geological Observatory, Palisades, N. Y.

proceed to obliterate all evidence of a sharply defined lithological break. In consequence it is usually impossible to detect such lacunae in a single core. However, if several cores from a particular region are available, a test of continuity can be applied. This involves making a layer-by-layer correlation between two or more cores. Presumably simultaneous slumping even at stations situated on two different topographical features is a possibility, but the probability that the same thickness of sediment will be lost from both is small. When the correlation includes three or more cores the probability of loss of the same thickness of sediment from all is infinitesimal, and continuity of the record may be regarded as fully proven.

In addition to climatic zones defined by dominance of either high or low latitude species of planktonic foraminifera, there are other zones that are defined by changes in percentage of dextral and sinistral tests of certain species of planktonic foraminifera. For correlation the most useful species is *Globorotalia truncatulinoides*. Examination of top samples of cores from the North Atlantic has shown that the areal distribution of this widely ranging species can be subdivided into provinces distinguished by dominance of dextral or sinistral tests (Ericson *et al.*, 1954). Just what the controlling factor in the environment may be is not clear. Apparently water temperature plays no part. However, from the core records, it is apparent that short period shifts in coiling dominance accompany abrupt changes in climate, as indicated by the ratio of high to low latitude species of foraminifera. At such times the normal pattern of distribution of dextral and sinistral populations must undergo a drastic rearrangement; soon after that the pattern reverts to normal again, regardless of whether the ensuing climate is of glacial or interglacial character. In any case, these short-period changes in coiling dominance are invaluable aids in the cross-correlation of suites of cores.

FIGURE 1 shows the sequence of six faunal zones the reality of which has been established by cross-correlation of suites of cores. These zones have been traced southward from the Canary Islands, across the equatorial Atlantic, through the Caribbean and Gulf of Mexico into the western Atlantic almost to the Azores. In order to emphasize the objective existence of these faunal zones in contrast to the necessarily subjective interpretation in terms of Pleistocene climatic events, I have designated them by letters. However, there is at least one faunal change that can be fitted into the climatic sequence with confidence, thanks to radiocarbon dating. The fact that the faunal change from zone y to zone x took place about 11,000 years ago (Ericson and Wollin, 1956) shows that it records the climatic amelioration that led to recession of the continental glaciers at the close of the last ice age. This conclusion is of importance particularly because it indicates what kind of faunal changes should be expected at earlier times of similar climatic change.

According to the almost universally accepted Pleistocene chronology of Cesare Emiliani, based on oxygen isotope paleotemperatures (Emiliani, 1955), the "Early Glaciation" or Nebraskan and the Antepenultimate or Kansan glaciations are included within zone v. However, no faunal change comparable to that at the boundary between zones y and z occurs anywhere within zone v.

In view of this, I present the following two alternative correlations for consideration.

As a first surmise I suggest that zone *x* corresponds to the warm interstadial of the last glaciation. Zone *w* then becomes equivalent to the early Würm glaciation, zone *v* to the last interglacial, and zone *u* to the final part of the penultimate glaciation.

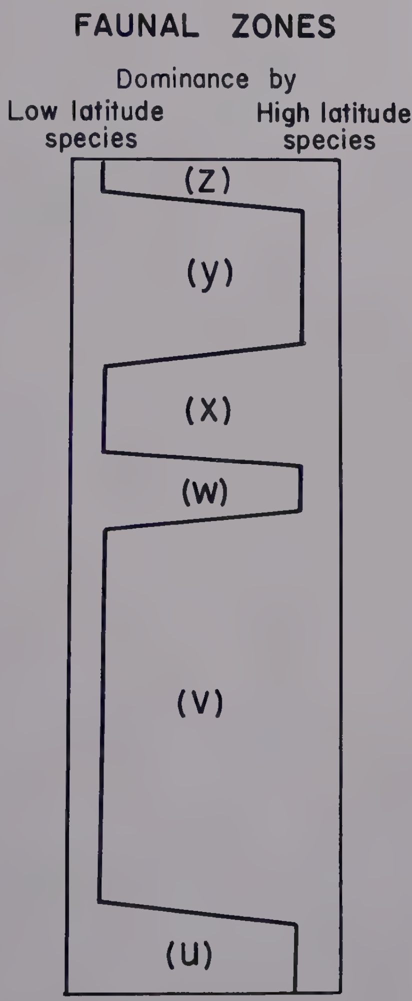


FIGURE 1. Normal sequence of faunal zones distinguished by dominance of high or low latitude species of planktonic foraminifera in sediment cores from the Atlantic, Caribbean, and Gulf of Mexico.

As a second possibility we may regard zone *x* as representing the last interglacial or Sangamon. Now zone *w* must represent the penultimate or Illinoian ice age, zone *v* the Great Interglacial or Yarmouth, and zone *u* the antepenultimate or Kansan ice age. A serious objection to this interpretation is that it forces us to place the warm interstadial within zone *y* where there is nowhere any indication of an important climatic amelioration. The slight amelioration

of climate suggested by cessation of turbidity current deposition in the Sigsbee deep cores is excluded from correlation with the warm interstadial by its date, 20,000 years B.P. For this reason I favor the first interpretation that makes zone x equivalent to the Würm 1/2 or warm interstadial.

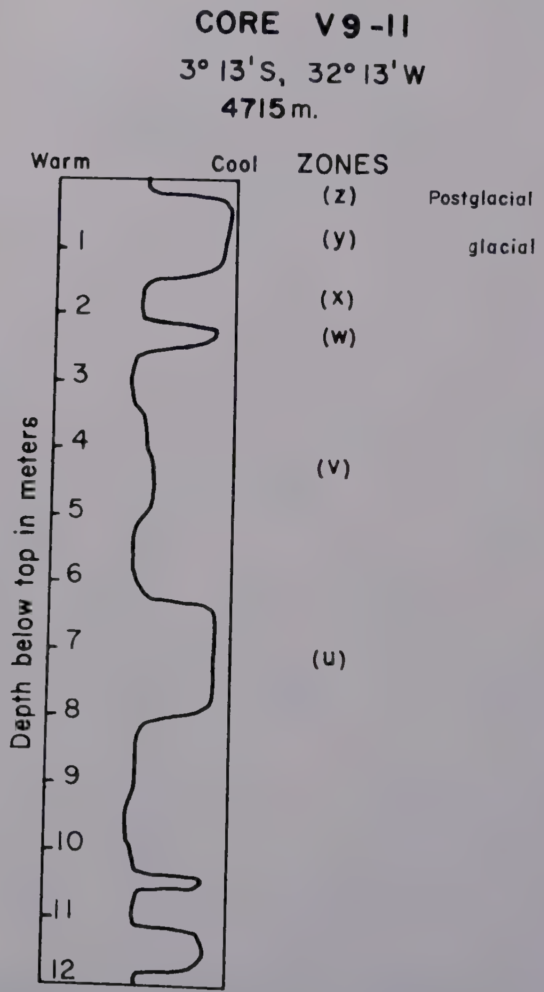


FIGURE 2. Climatic record in a sediment core from the equatorial Atlantic. The climatic oscillations are inferred from variations in relative abundance of certain species of planktonic foraminifera in samples taken at 10 cm. intervals.

FIGURE 2 shows the faunal zones in an exceptionally long core from the equatorial Atlantic. Since no other core reaches the zones below zone u, the lower part of the section can not be tested for continuity by cross-correlation. Admittedly the section below zone u may be incomplete because of slumping, but in that case the lower faunal zones are even older than extrapolation of the

rate of sediment accumulation would indicate. On the assumption that the section represents continuous sediment accumulation, the time represented is on the order of 600,000 years. According to my preferred interpretation of the faunal zones, the lowest zone of high-latitude species corresponds to the Kansan or antepenultimate glaciation. If we assume that zones equivalent to the First Interglacial and the Early Glaciation lie below, and that they are of about the same thicknesses as the overlying zones, we must conclude that the Early Glaciation or Nebraskan glaciation began about 1,000,000 years ago. If, on the other hand, we adhere to the short chronology of Emiliani (1955), zone *u* and the still lower zones of high-latitude foraminifera must correspond to phases of the Donau glaciation.

In any case it is my firm conviction that, contrary to prevalent opinion, the chronology of Pleistocene climatic events and the total duration of the Pleistocene epoch are still unsettled questions. I am confident, however, that the answers will be found eventually in longer deep-sea sediment cores.

References

Emiliani, C. 1955. Pleistocene temperatures. J. Geol. **63:** 538–578.

Ericson, D. B. & G. Wollin. 1956. Micropaleontological and isotopic determinations of Pleistocene climate. Micropaleontol. **2:** 257–270.

Ericson, D. B., G. Wollin & J. Wollin. 1954. Coiling direction of *Globorotalia truncatulinoides* in deep-sea cores. Deep-Sea Research. **2:** 152–158.

Ewing, M., D. B. Ericson & B. C. Heezen. 1958. Sediments and topography of the Gulf of Mexico. *In* Habitat of Oil. L. Weeks, Ed. Am. Assoc. Petroleum Geol.

CONVERGENCE OF EVIDENCE ON CLIMATIC CHANGE AND ICE AGES

Rhodes W. Fairbridge
Department of Geology, Columbia University, New York, N.Y.

Introduction

Consideration of ideal meteorologic circulation patterns over an almost "smooth" globe of medium-depth seas and low continents of moderate dimensions indicates that the thermal balance would prohibit "ice ages." The efficiency of upper-atmospheric heat transfer is assisted by the more sluggish but stabilizing control of oceanographic currents. Even with the known perturbations of the solar spectrum, variations in the cosmic-ray emissions, and observed astronomic-mechanical controls, the geologically "normal" climatic history of the globe shows no evidence of ice ages through the better-known range of the last 10^9 years except at certain specific times and places. Roughly speaking, those times were:

- Late Precambrian, 6 to 7 $\times$ 10^8 years ago. Place: Northern Hemisphere, Australia, South America, and South Africa; but not necessarily synchronously.
- Siluro-Devonian 3 to 4 $\times$ 10^8 years ago. Place: South America and South Africa.
- Permocarboniferous 2.8 $\times$ 10^8 years ago. Place: South America, South and central Africa, India, and Australia.
- Quaternary (Pleistocene) 5 $\times$ 10^5 years ago and still continuing. Place: North America, Greenland, Scandinavia, and Antarctica.

It may be noted that, from the present (limited) data in Antarctica, there is no positive evidence for any ice ages prior to the Quaternary (Fairbridge, 1952), except for a recent report of possible Permian tillites in the Horlick Mountains. While the age ranges given are rather broad, the actual duration of glacial phases within any particular period is quite brief, probably not exceeding 0.5 to 1 $\times$ 10^5 years in any one instance.

The recurrence of perhaps 50 to 100 such brief phases in 4 concentrated groups over a time span of 1 $\times$ 10^9 years suggests an "accidental" development or coincidence under an extremely low probability coefficient. In this paper it is submitted that ice ages are normal and predictable events under a long term sequence of observed variables in the physical history of the earth. The term accident is therefore intentionally used to indicate the rarity of certain events.

Prior to the Cambrian, both in the late Precambrian and older than 1 $\times$ 10^9 years, there is widely scattered but poorly known evidence of glacial formations from every continent. There is no evidence that the glaciation was synchronous, so that its distribution could be explained by polar movements, numerous orogenies, and possible crustal displacements. However, the contemporary pattern of geochemical reasoning about the original form and early evolution of the earth would ascribe a lowered temperature to early geologic time.

The idea of a cool accretion from a low-temperature dust cloud (ca. 50° K)

has been developed by Weizsäcker (1943), Hoyle (1955), Urey (1952), and Kuiper (1957). Although development and liberation of gravitational energy led to melting and differentiation, the present temperature distribution is anomalously low with respect to a calculated melting-point gradient (Ringwood, 1960), and implies considerable cooling, the heat transfer being both by convection and radiation.

A second thermal unbalance is proposed by Schwarzschild (1958), although in the short period of observational records the solar "constant" has been remarkably stable. In very extended terms, considering the evolution of stars, Schwarzschild submits that the sun's luminosity has increased by a factor of 1.6 since the origin of the earth (4.5 to 4.7×10^9 years). Ringwood (1961) calculated that this would lead to a rise in the earth's mean surface temperature by 30 to 40° C., which would imply that, prior to 3×10^9 years, mean temperatures would be below 0° C., and large areas of the earth would be glaciated. Very few sedimentary rocks are known that are older than this time, although one (3.7×10^9 years) conglomerate of uncertain climatic facies has been reported from the northern part of the Union of Soviet Socialist Republics (Lawrence Kulp, personal communication). Öpik (1953), who considers that variations in thermal luminosity in excess of 10 per cent are possible, would explain all ice ages by this mechanism.

The geological evidence is thus inconclusive as to whether the dominant pattern of these two opposite trends is cooling, warming or, indeed, oscillating. While violent oscillatory solar emissions of very short period are recognized, nevertheless the mean level of radiation must be very close to constant. This is proven by the principle of biological continuity. Considering that the surface temperature of a given planet could range from somewhere near 0° K to several thousand degrees, it is significant that normal vigorous metabolism is restricted to a very narrow range, about 0 to 50° C. Numbers of "mineralogic thermometers" exist, for example calcite/aragonite stability and gypsum/anhydrite stability, which confirm that mean ocean temperatures could not have departed far from 25 to 30° C. since 1×10^9 years ago. $CaCO_3$ secreting algae (*Collenia* types) are being discovered in progressively older Precambrian formations, putting the limit of climatic stability proved by biological continuity back to nearly 3×10^9 years.

Geologic Tectonic Revolutions

Consideration of the geotectonic evolution of the globe demonstrates that certain belts of the earth's crust have become consolidated within specific periods of time. Each tectonic event is known as a phase, and a group of phases constitutes a revolution, marked by a definite sequence of crustal movements, sedimentation, magmatic activity, and orogenesis (Stille, 1924). It has been pointed out by certain geologists that, in particular highly mobile regions such as the Californian coast ranges, the number of such phases seems almost infinite and the duration of the revolution so long as to destroy the concept of periodicity (Gilully, 1949). However this is not true for the whole earth, which shows a marked stability after the termination of a specific revolution and permits the construction of geotectonic maps delimiting such regions

in time and space (Umbgrove, 1947). One may note that currently a subcommission of the International Geological Congress is preparing a detailed world geotectonic map that will illustrate this fact.

The concept of orogenic periodicity is well established, although not with the sharpness or exclusive limitations perhaps implied by Stille. Vulcanicity and orogeny go hand in hand, so that the great revolutions are therefore marked, not only by the production of high mountains but also by the effusion of large volumes of volcanic ash and dust.

Between such mountain-building revolutionary episodes, extended periods of crustal quiescence are known. Only locally, in restricted belts, were these quiescent periods disturbed by further mountain building. Thus the orographic appearance of the earth varied progressively through stages of extreme relief to extreme flatness. One may cite the Ordovician and the Cretaceous as periods of particular quiescence, although both closed with revolutionary tectonic events. The distribution of thin, flat-lying sediments of each of these periods over very broad areas of the stable parts of the globe testifies that sea level was eustatically high; such stages are known as thalassocratic. In contrast, times of excessively low sea level were engendered by such factors as crustal folding and the formation of deep sea trenches; the sea then flowed back into these depressions and the effective land areas became enlarged, thus *epeirocratic*. Judging from statistical analysis of the present earth's topography and considerations of the former distribution of parageosynclines and epeiric seas (Fairbridge, 1955, 1959), it would seem that only the thalassocratons (stable deep-sea crust) would remain permanently submerged, intermediate areas being geotectonically reversible. Of the present total surface of the earth, 510×10^6 km.^{2}, about 361.9×10^6 km.^{2} are covered by ocean; 148.1 by land (see FIGURE 1). A general retreat, an epeirocratic phase, would expand the land area to possibly 180×10^6 km.^{2}, while widespread peneplanation and the resultant thalassocratic phase would reduce the land area to only about 110×10^6 km.^{2} In such oscillations both tectonoeustasy and glacioeustasy may play a part.

The meteorologic implications of such areal changes are obvious. The paleoclimatic data for each thalassocratic stage evidenced by widespread coral reefs, warmth-loving plants, vigorous reptile evolution, and similar phenomena confirm the theoretically derived concept of general mild-warm maritime climates during much of the geological past. The geologically "normal" climate was subtropical. For the epeirocratic stages, extreme continentality is confirmed by evidence of deserts, red beds, ephemeral lakes and—not infrequently—glaciation. Epeirocratic epochs were geologically rarer than thalassocratic phases which, with their low relief and decreased circulation, were geophysically more stable. However, an expanding earth (Egyed, 1956, 1961), if the volume of ocean water remained approximately constant, would entail a progressive increase in the epeirocratic character.

Polar Topographic Coincidence

The phenomenon of polar migration has long been suspect on stratigraphic and paleontologic grounds, but the concept is now greatly strengthened by the

widespread paleomagnetic surveys of Runcorn (1956, 1959; Collinson and Runcorn, 1960), Irving (1956), Nairn (1960), and others. There is an internal consistency of their observations within any one continent, but appreciable differences between continents suggest not only some continental displacement but, quite likely, a certain element of mantle expansion (Egyed, 1956, 1961; Carey, 1958; and Heezen, 1960).

It was long ago recognized by Charles Lyell that changes of land and sea would effect climatic changes (see Schwarzbach, 1953, 1961). Equatorial continents would raise world temperatures, and polar continents would lower them. Wilhelm Ramsay (1924) evolved further the relief hypothesis, which coupled high relief (orogenic periods) with lowered world temperatures (miothermals), and low relief (thalassocracy) with milder world conditions (pliothermals). Favorable topography has become an integral part of most modern

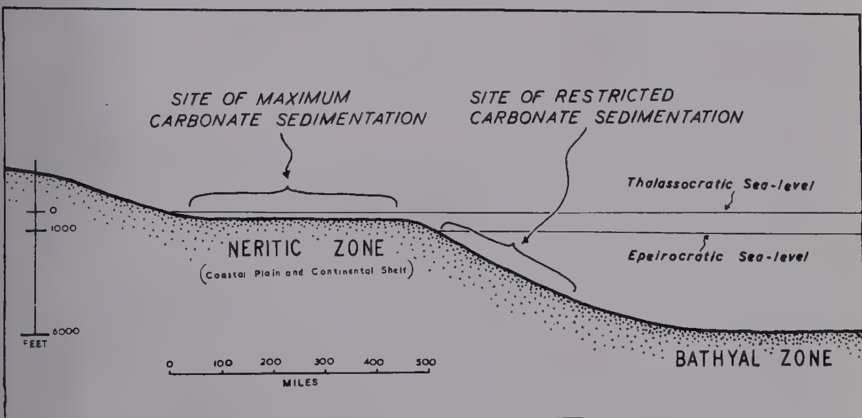


FIGURE 1. Effect of eustatic changes of sea level on continental lands. Epeirocratic phases coincide with and help to amplify ice-age cooling. Thalassocratic (ocean-dominated) phases correspond to interglacial and nonglacial times, the widespread ocean circulation effecting a general amelioration of world climates.

ideas in paleoclimatic change (for example, the solar-topographic hypothesis of Flint, 1957).

As pointed out by Köppen and Wegener (1924), the migration of the poles into the present arctic- antarctic topographic pattern in late Tertiary times would profoundly affect the meteorologic circulation. Milankovitch (1930, 1941) readily conceded that his solar-radiation theory would be relatively ineffectual without an appropriate geographic setting. However, the early idea that rapid Quaternary polar migration would explain the glacial/interglacial cyclicity is now completely rejected (Zeuner, 1959).

A progressive late Tertiary cooling of the temperate belts in both Northern and Southern Hemispheres is shown by paleobotanical and oxygen-isotope evidence from Oligocene to late Pliocene. Data are available on a world-wide basis from: the western United States (Barghoorn, 1953; Durham, 1959; Dorf, 1959); Europe (Woldstedt, 1954; Teichmüller, 1958); Siberia (Zaklinskaya, 1958); eastern Asia (Chaney, 1940); Antarctica, that is, Grahamland (Hennig,

1911); Australia (Dorman and Gill, 1959*a* and *b*); and both Atlantic and Pacific Ocean deep-sea sediments (Emiliani, 1954, 1955, and 1961). Progressive drop of sea level through this late-Tertiary period suggests a build up of mountain glaciers, but no large icecaps are known. Former, pre-Quaternary ice ages could be initiated in like manner, following great world-wide orogenic events.

At the same time mountainous Antarctica, a continent of about 5 million square miles (14×10^6 km.^{2}), came to be symmetrical about the South Pole. Stratigraphic evidence suggests gradual cooling in West Antarctica until the Pliocene and probably mountain glaciation had set in. A permanent icecap formed only in the late Pliocene or early Pleistocene, and it still persists today, causing and maintaining a condition of unusually low sea level (Fairbridge, 1952).

Glaciation was also initiated in Greenland at least by Pliocene times (Wager, 1933). According to Emiliani (1961), ocean-bottom temperatures had reached present levels by late Pliocene, confirming these data; the total drop since Upper Cretaceous was 12° C.

As Ewing and Donn (1956, 1958) noted, following modern paleomagnetic surveys, this migration of the poles in late Tertiary times in the Northern Hemisphere brought the North Pole into the nearly landlocked Arctic basin, an event that would have considerable reactions in the waters of the Arctic Ocean; such a condition must further lead to a modification of the Arctic-Atlantic circulation and then to a world cooling. In my view, in contrast to that of Ewing and Donn, the initiation of mountain glaciation in Antarctica was far more important than the cooling of the Arctic Sea (which would have no eustatic effect by itself). The weakness of this polar-migration event, by itself, is that it is noncyclic, but eustatic oscillations over the Nansen/Iceland sill will undoubtedly operate as a shutter mechanism, which is the key to the Ewing and Donn hypothesis.

However, the early Tertiary was a time of thalassocratic sea levels and uninterrupted east-west equatorial currents (see FIGURE 2; and Berry, 1922) but, from late Tertiary onward, there has been a gradual swing to an epeirocratic condition (Chamberlin, 1899), accompanied by a reorientation of oceanographic circulation to North-South directions and restricted cells (see FIGURE 3). Furthermore, the increasing salinity and reduced ocean areas would lower evaporation rates (Visher, 1921). Our observations regarding the eustatic oscillations (Fairbridge, 1960, 1961) suggest that prior to the *last* glacial (Wisconsin), the Nansen/Iceland barrier would be progressively less and less effective in the earlier stages because sea level in the Pliocene was probably 200 m. above the present; since then there has been a secular drop in sea level that has been independent of glacioeustatic oscillations (Zeuner, 1959). Moreover, the timing of the shutter mechanism presents serious difficulties as an explanation of climatic cyclicity. The key to ice ages in general, in my opinion, is therefore what I might call the "polar coincidence theory." In the case of the Quaternary ice age it is essentially "antarctic coincidence."

Eustatic-temperature control. It is generally said that in high latitudes the temperature drops approximately 1° C. for every 100 m. of elevation; thus a

glacioeustatic lowering of this order caused by the glaciation of Antarctica would by itself go a long way toward promoting a world-wide cooling (Zeuner, 1959). The mean air-temperature curve at sea level, if lowered 1° C. at lat. 45°, would drop off exponentially so that at lat. 80° the temperature would be lowered by the order of 10° C.

Tectonoeustatic cooling. As noted above, there was an additional, progressive tectonoeustatic drop of the order of 50 to 100 m. through the late Tertiary/Pleistocene due to certain geotectonic processes: the lowering of the Melanesian-East Indian-Mediterranean-Caribbean basins (Fairbridge, 1961). The idea of subsidence control was stressed already in the last century by Geikie (1894). The combined glacial and tectonic curve, as now visualized (see FIGURE 10), is an oscillation of amplitude about 100 m., superimposed on a secular slope of about 100 m. in 200,000 to 600,000 years, flattening during the last 50,000 years and, possibly, reversing (reflecting geotectonic stabilization). The temperature lowering in the Wisconsin* from the interglacial maximum might cool the 80° N lat. coastal stations by 20° C. at sea level (see FIGURE 9). Relatively small thermal changes at the equator are magnified poleward, owing to the progressive reduction of unit areas in higher latitudes (Schell, personal communication).

The progressive lowering of sea level throughout the Pleistocene (attributed by Zeuner and myself to geotectonic causes) might be alternatively interpreted as due to a combined over-all cooling, in addition to the observed glacial oscillations. The O^{18} isotope data from warm Atlantic deep-sea cores by Emiliani (1955) suggests that there was indeed a marked oscillation (5 to 8° C.). However, there was no secular temperature lowering from the beginning to the end of the Pleistocene. Nevertheless, there was a general lowering of bottom temperatures from middle Tertiary times onward. The evidence against a wholesale melting of the antarctic icecap with each interglacial suggests that with each new glacial onset there may have been a progressive augmentation of the total ice mass. Thus a certain fraction (for example, 50 m.) of the post-Pliocene 200 m. lowering, attributed here to both glacial and geotectonic causes, may be due to progressive noncyclic glacial accumulation. At the present time we do not appear to have any evidence as to the dimensions involved.

Long-period variations in solar radiation. As noted by Haurwitz (1946) and others, there is little observed variation in the solar constant in the present century, although there is an appreciable variation in the UV light component. Many observers, however, have ignored the fact that the present century is one of general thermal advance. In contrast there are well documented periods of general thermal lowering, which coincide precisely with times of very few sunspots. Current means oscillate in a 22- and 40-year swing between 20 to 30 spots at lows, and 50 to 60 spots at highs (Willett, 1951) while, in the late 17th century, the means were down to 5 or 10 spots. That climatic events coincide with the sunspot cyclicity has been claimed for a long time (see, for

* While there is a certain temptation to regard the major advances in North America, as generated by climatic changes of the same magnitude, data from Europe (Woldstedt, 1954, 1958) and the Soviet Union (Kats, 1960), suggest that the last glaciation was not so great as earlier advances.

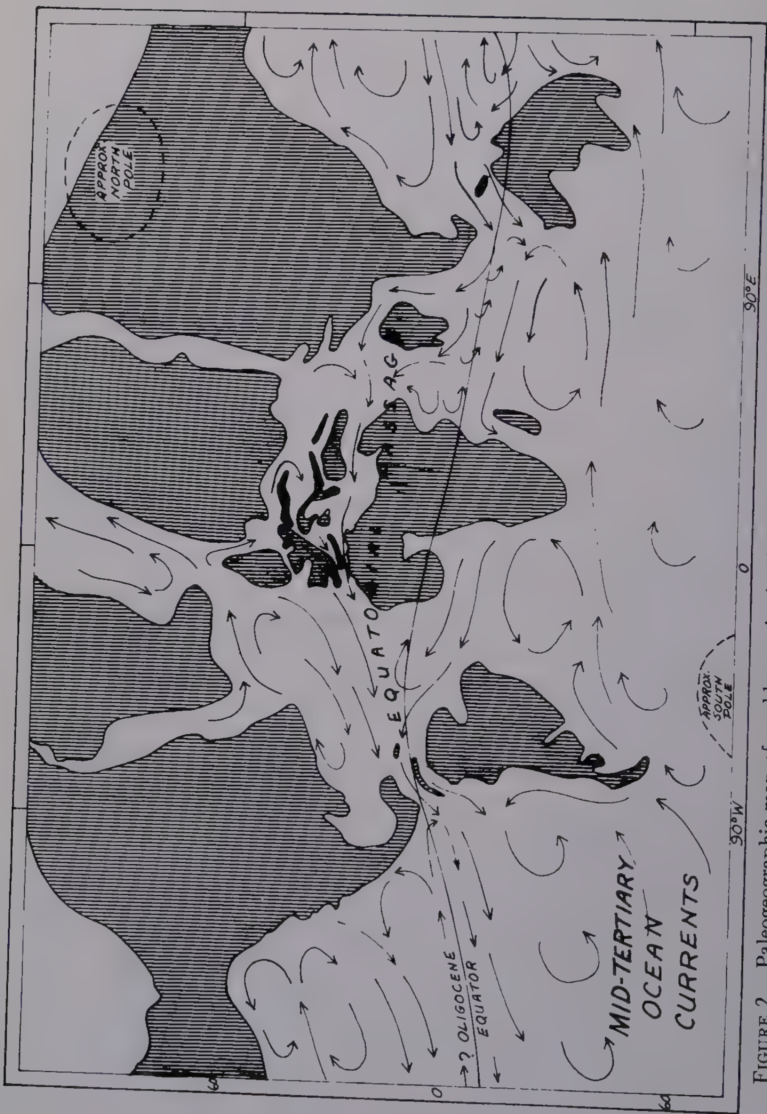


FIGURE 2. Paleogeographic map of world oceanic circulation during the middle-Tertiary (approximately 30 million years ago). Note the nonexistence of the present north-south barriers and the world-encircling "equatorial passage." Temperatures at the floor of the tropical deep seas were at least 8° C. warmer than today, and high-latitude surface temperatures were warmer still (Emiliani, 1961).

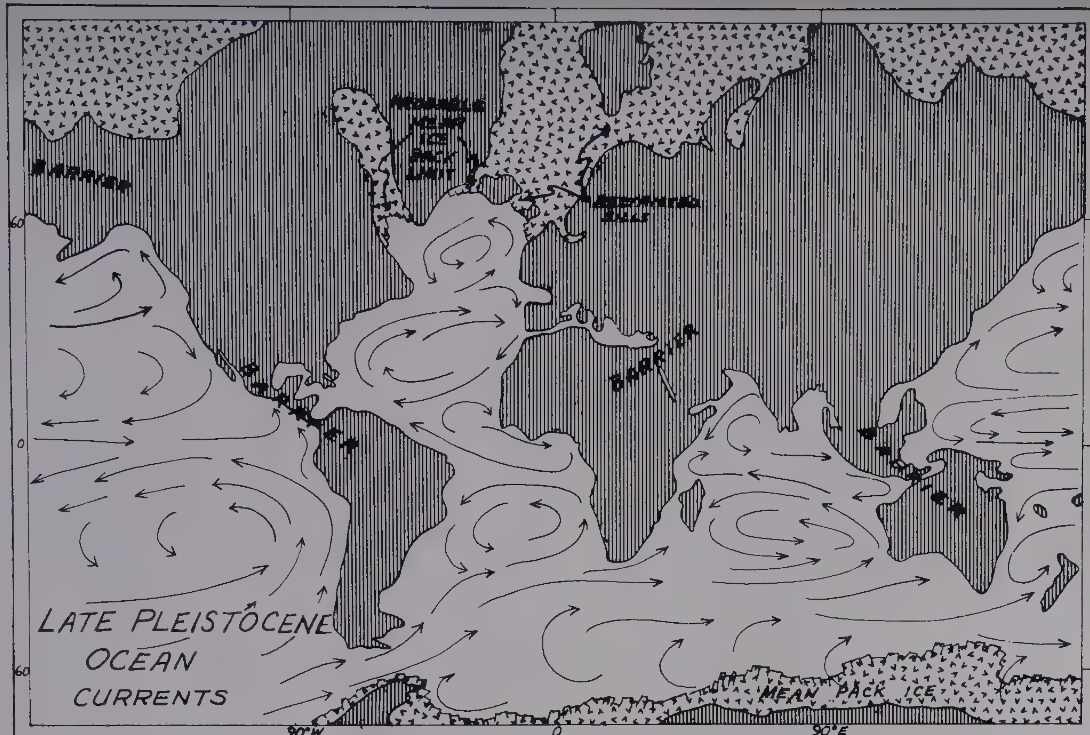


FIGURE 3. Paleogeographic map of world oceanic circulation during the late Pleistocene (last glaciation, about 20,000 years ago). Note the appearance, due to geotectonic action, of the Isthmus of Panama, a narrow Strait of Gibraltar, a total closing of the Aleppo platform in the Middle East, blocking of the Tethys Sea across northern India, narrowing of the East Indian seaways, and closing of the Torres Strait Islands. The appearance of restricted circulation gyres and north-south cold water transfer contrasts strikingly with the warm east-west systems of FIGURE 2.

example, Clayton, 1920; Anderson, 1939; and Abbot, 1957), but the nature of the relationship is not at all clear. The global circulation is clearly involved (Köppen, 1914; Flohn, 1952).

As Willet (1949) showed by instrumental analyses, there are regional reactions to the growth of sunspots (in opposing signs), which confused the picture. Up to a certain sunspot level, there may be a temperature rise, but then a counter oscillation intervenes and a reversal occurs. Thus the climatic reactions to sunspot maxima (150 to 200 spots) often differ notably from moderate peaks (50 to 100 spots). One may recall that Simpson (1940, 1957) pointed out that, in a general solar radiation cycle, maximum precipitation should occur on the slopes of the curve, the solar minima calling for long dry "interglacials" and the maxima with short wet ones. There seems no reason why the same principle should not apply to brief solar-derived climatic reactions, although Simpson's oversimplified explanation of the over-all Pleistocene history must be rejected (Schwarzbach, 1961).

It seems likely that, superimposed upon a general increase in orographic cold, eustatic lowering, and so on, a succession of such cycles of low sunspot activity could have materially contributed to the Pleistocene cold phases. Successions of even relatively short low sunspot activity would help to establish mountain-glacial conditions in the higher latitudes and then, in turn, by lowering sea-level and other sequential effects, would help initiate the cooling chain reaction (a feedback process).

To identify long-term variations in solar radiation from geologic data is not a simple matter. First, one should recognize the difference between actual dynamic radiation variability (evidenced by the sunspot cycle, cosmic-ray showers, and other factors) and effective or received radiation (controlled by "mechanical" factors, such as the earth's astronomic position and attitude).

In the first category, dynamic variability, there is the meteorologic evidence that there is a coincidence, admittedly not at all well understood, between sunspot periodicity and climate. Generally speaking, the situation is this: there is a cycle of about 11.15 years (Dewey, 1958) for sunspot numbers that range from 3 to 5 at the minima to 40 to 200 at maxima. The extraordinary range of the maxima is associated with greatly variable climatic reactions.

The 11-year "cycle" is the mean of a 9 to 13 year oscillation that tends to break up into a high-low alternation where the "highs" tend to be 20 to 50 spots above the low maxima; this is the "Hale cycle" of 22 to 23 years (22.75 years, according to Abbot, 1957), which coincides with the reversed magnetic polarity of the spots (Anderson, 1939). Also there is the "double Hale" cycle of 40 to 44 years; this may be related to waves in the 44-year oscillation of the geomagnetic field, which also shows a semiwave of 22 years and quarter wave of 11 years (Barta, 1956). One observes that the 11-year sunspot sequence tends to occur in 3 to 4 phase low series and 3 to 4 phase high series. This gives the basis for Schove's 77.5- to 78.0-year sunspot cycle and a 160- to 170-year cycle. Gleissberg (1958) believes that there is an 80-year astronomic cycle in this sequence, and he has plotted 20 such oscillations since 400 A.D.

Looking back over the rather imperfect sunspot record of the last two millenniums (FIGURE 4), derived by Schove (1955) from classical Chinese aurora records and checked by other historical data, the long-term variability becomes

much more apparent. However, the inadequacies of the documentation prevent rigorous cycle analysis.

To demonstrate that long-range variation in sunspots is reflected by worldwide climatic change is not easy, although variability in the meteorologic records for certain regions such as northwestern Europe during the last 500 years shows a reasonable match with the radiation curve (Manley, 1953; Baur, 1948; Labrijn, 1945; and Ångström, 1939). Variations in the levels of the Caspian Sea (Huntington, 1907; Berg, 1940; Butzer, 1958; and Fedorov and Skiba, 1960) and of the Great Salt Lake (Brooks, 1951) also reflect the sunspot record. A plot of the geological (C-14 dated) and historic record of eustatic changes over the last 2000 years shows an oscillation through a maximum amplitude of 2.5 m. A smoothing of the sunspot curve to 100-year means produces a striking similarity (see FIGURE 4).

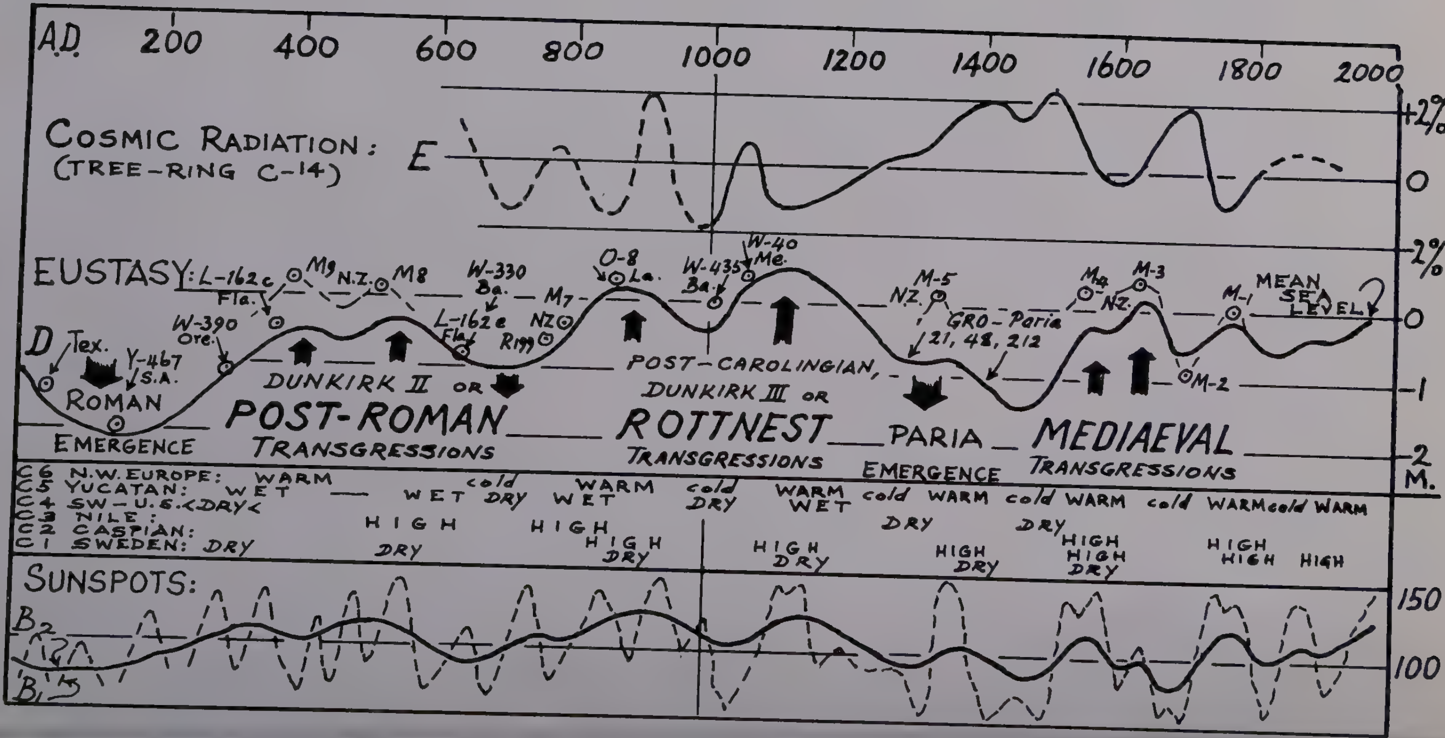
The mechanical relationship between climate and sunspots is striking, but causal connections are more difficult to establish and are beyond the scope of the present paper. Suffice it to say that at the present state of knowledge it appears that bursts of solar UV radiation, exceeding by 200 per cent the normal activity, coincide with sunspot maxima. The X rays also show this periodicity. The UV wave-length emissions generate the formation of ozone (O_3) from oxygen in the upper atmosphere, and ozone plays an important role in the earth's thermal blanket ("greenhouse effect"). The relationship is suggestive, but not yet quantitatively proved (Kraus, 1960; Freidman, 1961).

Other greenhouse-effect agents (CO_2 and H_2O vapor) appear to play contributory roles, although possibly overstated by some advocates (for example, Chamberlin, 1899; lately by Plass, 1959, 1961). Volcanic dust, following the theory of Humphrey (1913), certainly seems to be reflected statistically in world-temperature means (Abbot and Fowle, 1913; Wexler, 1952; Lamb and Johnson, 1959; and Mitchell, 1961), but the effects are highly transient, at least with current vulcanicity.

Geological data reflect still-longer variables. These chiefly seem to fall into a second category, celestial mechanics being responsible. The position or attitude of the earth with respect to the sun results in variations in the effective solar radiation ("insolation") that is received by latitudinally distinct unit areas of the earth. The asymmetry of the earth's topography excludes any dynamic balance.

An astronomic variation of about 550-1100-1650 years is known (earth/moon orbital motion). C. M. Stacey (unpublished) notes that maximum perigeal spring tides occur approximately every 1668 years; at this time the moon is in perigee, the earth at perihelion, which is Stacey's "zero check cycle" that had its last culmination at 1433 A.D. (517 B.P.). Pettersson (1914) first applied this to his "tide generating force," which he visualized as breaking up arctic ice floes. The one-half harmonic (1133 years) has been designated by Karlstrom (1961) as his "stadial cycle." The one-third harmonic (567 years) seems particularly frequent in geologic records (Pearson, 1901; Bakker, 1958; and Fairbridge, 1959). This 567-year period itself is one twelfth of the 20,400-year precession cycle, and one sixth of Karlstrom's "substage cycle" (ca. 3400 years).

In Scandinavia, peak years in the melting of the last continental ice sheet


A.D. 200 400 600 800 1000 1200 1400 1600 1800 2000
COSMIC RADIATION: (TREE-RING C-14)
E
+2%
0
-2%
EUSTASY: L-162c Fla.
M9 N.Z. M8
W-330 Ba.
M7
O-8 La.
W-435 Ba.
W-40 Me.
M-5 N.Z.
GRO - Paria 21, 48, 212
M4 N.Z. M-3
M-1
MEAN SEA LEVEL
M-2
D Tex.
Y-467 S.A.
W-390 Ore.
L-162e Fla.
NZ
Rigg
ROMAN EMERGENCE
DUNKIRK II OR POST-ROMAN TRANSGRESSIONS
POST-CAROLINGIAN, DUNKIRK III OR ROTTNEST TRANSGRESSIONS
PARIA EMERGENCE
MEDIAEVAL TRANSGRESSIONS
0
-1
-2
M.
C6 N.W. EUROPE:
C5 YUCATAN:
C4 SW-U.S.:
C3 NILE:
C2 CASPIAN:
C1 SWEDEN:
WARM
WET
DRY
cold
WET DRY
WARM
WET
cold
DRY
WARM
WET
cold
WARM
cold
WARM
cold
WARM cold WARM
HIGH
DRY
HIGH
HIGH DRY
HIGH DRY
HIGH DRY
DRY
HIGH DRY
DRY
HIGH HIGH DRY
HIGH HIGH HIGH
SUNSPOTS:
B2
B1
150
100

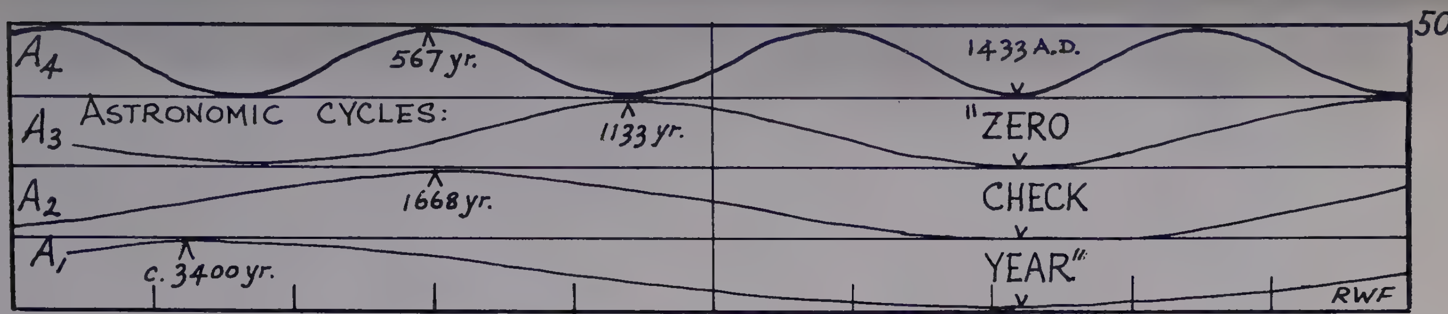


FIGURE 4. Eustatic, astronomic, radiation and climatic curves for the last 2000 years.

(A) Astronomic cycles: A1, Karlstrom's substage cycle of 3400 years; A2, Pettersson's tide cycle of 1668 years; A3, Karlstrom's stadial cycle of 1133 years; and A4, 1/3 harmonic of Tide Cycle—567 years. The "zero check year," 1433 A.D., is according to calculations by Stacey and by Karlstrom.

(B) Sunspot cycles: B1, 30-year smoothing of Schove's annual means; and B2, 100-year smoothing of the same.

(C) Selected climatic indicators: C1, Swedish "recurrence horizons" (dry, high-latitude, warm stage in the cool-wet-loving peat sequences) adapted from Granlund (1932) and von Post (1946); C2, records of Caspian Sea levels, reflecting middle-latitude evaporation and precipitation (Berg, 1940; Butzer, 1958; and others); C3, records of Nile flood levels (Brooks), reflecting subtropical rainfall (probably coincident with northern warmth, C1); C4, records of drought in the United States Southwest, reflecting high-pressure systems in middle latitudes and decreased meridional transfer (mainly converse to C1–3 and 5); C5, records of excessively wet periods in Yucatan (Central America) reflect increased tropical activity (should match C3); and C6, North-west European temperature characteristics (Brooks, and others).

(D) Eustatic sea levels based upon available radiocarbon dates (in this range, accurate mainly to about ±100 years) also upon stratigraphic, archeological, and historical records. Note similarity to 100-year smoothing of sunspot record. C-14 and other samples are indicated by laboratory numbers (see Fairbridge, 1961; and for N.Z. M1–9, see Schofield, 1960).

(E) Cosmic radiation based upon C-14 analyses of tree rings over the past 1300 years (Willis *et al.*, 1960; Stuiver, 1961), which display departures up to 2 per cent, with a periodicity that suggests an inverse proportion to sunspot activity.

are counted in the varved lake sediments at 10,500, 9950, 9400, and 8900 years B.P. (de Geer, 1959). The growth rate of peat swamps is conditioned also by the temperature and rainfall in Scandinavia and, approximately at the same interval, peat shows a dying out and desiccation, followed by "recurrence horizons" at 4250, 3150, 2550, 1550, and 750 (each ±100) years B.P. (Granlund, 1932; von Post, 1946). The flooding of the Netherlands through the last 3 millenniums is given a similar period by Bennema (1954), and the early historical records of similar flooding in many parts of Europe show this same pattern (Pearson, 1901). Readvances of the ice in the "little ice ages" of Alaska and elsewhere are compatible with the same evidence (Fairbridge, 1961; Karlstrom, 1961).

There are undoubtedly longer oscillations to be extracted from the geological record. To what extent these are due simply to aperiodic swings from one extreme to another, or to a complex cyclicity, is still a much-debated question. The most attractive hypothesis is the so-called Croll-Milankovitch-Bacsák theory of effective solar radiation. It is based on celestial mechanics, and is mathematically well founded on cycles of 21,000, 40,000, and 92,000 years (see Zeuner, 1959). Its climatic reactions and correlations are the factors that have caused the controversy. Before discussing this theory additional observational data will be summarized.

Quaternary temperatures. An exhaustive review of Quaternary climates is not pertinent here, but a few highlights will be mentioned. Valuable summaries have been prepared by Büdel (1951), Flohn (1953), Flint (1957), Butzer (1958), Durham (1959), Emiliani and Geiss (1959), Schwarzbach (1961), and others.

Paleobotanic evidence is very useful in this field of paleotemperature measurement. The method is essentially a simple one. Statistical counts are made of pollen, separated from sediment samples taken at fixed intervals from the stages in question; these indicate relative increase or decrease of significant species of trees, shrubs, and grasses that can be used as mean-temperature indicators. The classical example was presented by Gunnar Andersson (1902, 1919); in Sweden today the limiting isotherm of the hazel is 12° C., but its pollen distribution 6000 years ago extended as far north as today's 9.5° C. isotherm, and it is thus deduced that the climate was then 2.5° C. warmer than today. There are of course many complicating factors such as changes in wind regimes, growth periods, sea level, and isostasy, but the use of multiple species and statistical analysis apparently confirms the original generalization (von Post, 1946).

Thus at the 3-m.-high sea-level stand of the "Atlantic" ("Older Peron") climatic stage (5000 years B.P.), the mean temperatures of northern Europe were 2.5° C. higher than today (see FIGURE 5). During the 2- to 3-m.-low sea level of the "subboreal" ("Pelham Bay") climatic stage (3000 years B.P.) the paleobotanical data indicate mean temperatures about 1.5 to 2° C. below the present. It is apparent, therefore, that if the long-period (550 years) cycle evidence is any criterion, a 0.5° C. climatic change is matched by about 800 mm. change in sea level. This is nearly an order of magnitude larger than that for the short-period cycles (80 to 160 years). Furthermore, if one considers

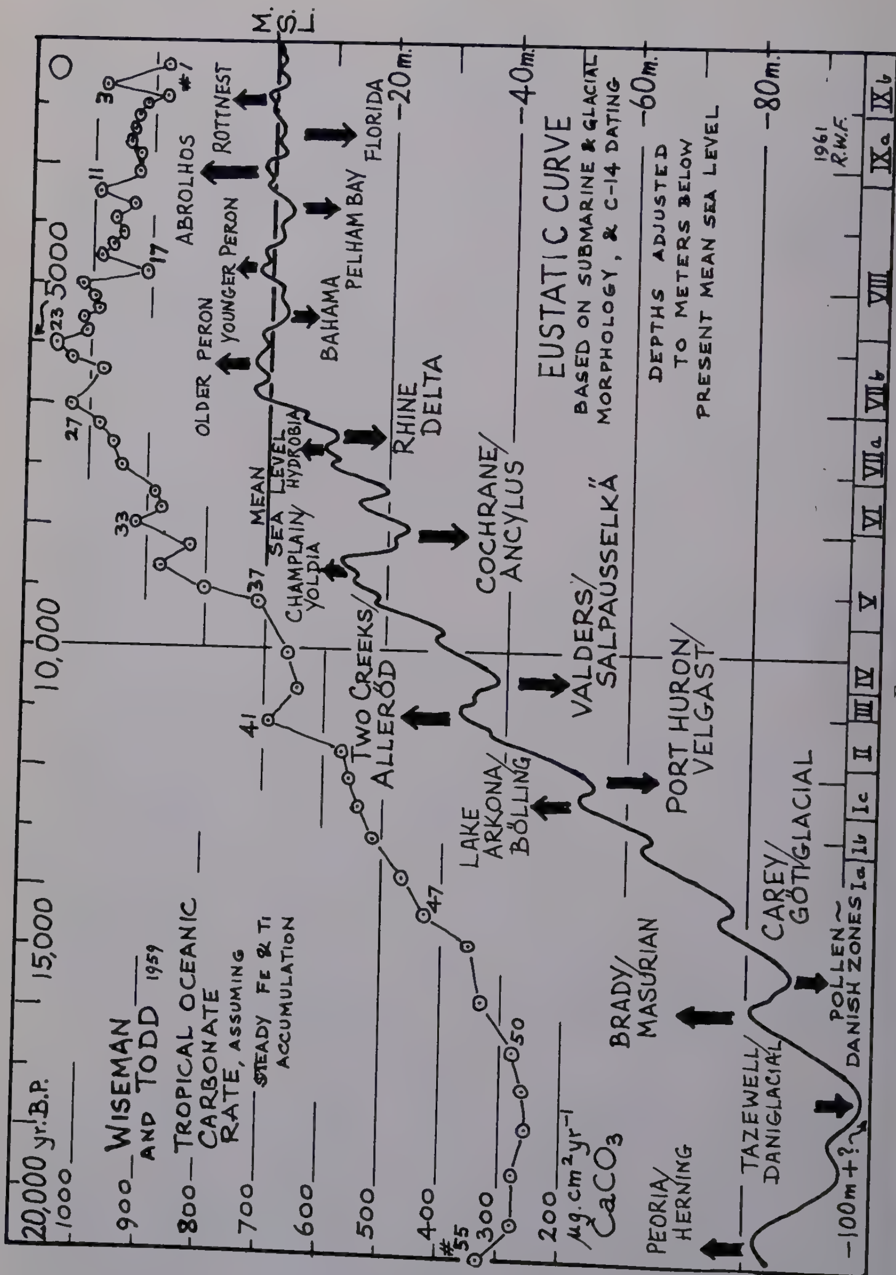
the very short cycles (5 to 11 to 22 years), there are sea-level oscillations, apparently eustatic, of the order of 80 mm. to 1° C.

Turning now to the still greater variations of the Quaternary glacial stages there are both eustatic data and paleoclimatic data from several sources that are reasonably compatible. The paleobotanical evidence from western Europe suggests that during the glacial maxima the temperature was 6 to 7° C. below the present and, during interglacials, up to about 2 to 3° C. above the present (Brooks, 1949, 1951; Manley, 1953; and Flint, 1957, p. 487). For the tropical belts a drop from 27 to 23° C., a change of approximately 4° C., is suggested by Flohn (1952). A cooling of the same order (4° C.) is implied from the 300 m. depression of the firn line in the glacials (Klute, 1930). In eastern Africa both firn and tree lines reached 800 m. lower (Klute, 1935). However, in the equatorial rain forest itself, Büdel (1959) considered that the cooling would only have been about 2° C. Confirmation of these data comes from associated animal fossils, and by O^{18} isotope determinations from the pelagic organisms in the warmer parts of the oceans (Emiliani, 1955), which showed a 5 to 8° C. amplitude oscillation from glacials to interglacials. Evaporation in the middle and low latitudes during the glacials was about 20 per cent lower than during the interglacials, owing to the lower mean temperatures (Flohn, 1953), so that when the jet stream was displaced about 10° of latitude toward the equator and the cool pluvial belts about 20° to the south, there was increased rainfall in Mediterranean belts (see FIGURE 6).

On the basis of contemporary instrumental analyses, Kraus (1955, 1960) has been able to show that a major cooling trend associated with a strengthening of the general circulation, with an expansion of the circumpolar vortex, would lead to a narrowing of the subtropical high-pressure belts (normally the "arid zones") and a narrowing of the equatorial (high-rainfall) belt. Kraus (1956) notes an inverse relation between the strength of the zonal circulation ("high index") and the monsoon-type circulations. Indeed the extended ice cover in the Arctic Ocean (4 per cent of the world's oceans), the large continental ice sheets (26 per cent of the world's land area), and the eustatic reduction of oceanic area (by about 5 per cent) must have further reduced evaporation; the present oceanic evaporation represents 85.8 per cent of the world total (Lvovich, 1961). Thus there can be no question of a glacial stage being a high-rainfall or pluvial period of world-wide character, in contrast to the views of Büdel (1959) and others (see discussion in Zeuner, 1953).

Sensitive Latitudes

Glaciological researches in Greenland and Antarctica over the last few decades have shown that the hydrologic balance of these last two great continental ice sheets is essentially in equilibrium. This is in contrast to the earlier views expressed, for example, by Hobbs (1911) that they are slowly melting. Certain of the recent determinations indicate a slight positive balance: that is, snow is building up faster than it is decreasing by calving, melting, ablation, and similar processes. The eustatic evidence permits no other interpretation. While northern Greenland shows a systematic decrease of precipitation over the period 1920 to 1957 (Diamond, 1958), the margin of east Antarctica shows an excess



20,000 yr. B.P.
15,000
10,000
5000
0
WISEMAN AND TODD 1959
TROPICAL OCEANIC CARBONATE RATE, ASSUMING STEADY Fe & Ti ACCUMULATION
μg. cm² yr⁻¹ CaCO₃
1000
900
800
700
600
500
400
300
200
M. S. L.
MEAN SEA LEVEL
-20m.
-40m.
-60m.
-80m.
-100m+?
EUSTATIC CURVE
BASED ON SUBMARINE & GLACIAL MORPHOLOGY, & C-14 DATING
DEPTHS ADJUSTED TO METERS BELOW PRESENT MEAN SEA LEVEL
PEORIA/HERNING
TAZEWELL/DANIGLACIAL
BRADY/MASURIAN
CAREY/GÖTIGLACIAL
POLLEN~DANISH ZONES
LAKE ARKONA/BÖLLING
PORT HURON/VELGAST
TWO CREEKS/ALLERÖD
VALDERS/SALPAUSSELKÄ
COCHRANE/ANCYLUS
CHAMPLAIN/YOLDIA
HYDROBIA
RHINE DELTA
OLDER PERON
YOUNGER PERON
BAHAMA
PELHAM BAY
ABROLHOS
ROTTNEST
FLORIDA
1961 R.W.F.
Ia Ib Ic II III IV V VI VIIa VIIb VIII IXa IXb

FIGURE 5.

of snow accumulation over ablation (Loewe, 1956; Crohn, 1959; and Mellor, 1959).

Such calculations may be accurate only for limited areas, and extrapolation can lead only to error, since the world mean sea level is essentially stable and has been so for 6000 years. The recent 1.2 mm. per-year rise of sea level is a quite minor feature, and is a normal aspect of the short-period, low-amplitude climatic oscillations of the Holocene epoch (Flohn, 1958).

Considering the eustatic record of the last 16,000 years, it is thus apparent that a major threshold was reached soon after 6000 B.P. A certain melting tendency was then completed and, for the last 5000 years, a slightly negative eustatic trend has been in progress. An inspection of the available ice masses shows that 6000 B.P. must have marked the end of the final phase of the melting

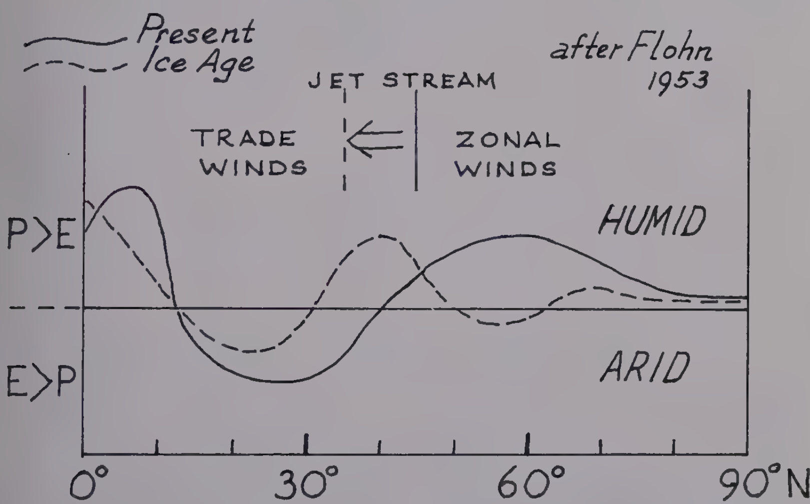


FIGURE 6. Diagram showing Quaternary glacial/interglacial displacement of humid and arid climatic belts. Note the equatorward displacement of the jet stream by 10° during glacials (adapted from Flohn, 1953). Also it should be stressed that the present interglacial (late Holocene) is only intermediate and not so warm as earlier thermal maxima. One may visualize a high interglacial displacement to the right. P = precipitation; E = evaporation.

of the North American-Scandinavian-Siberian ice sheets (de Geer, 1959; Henderson, 1959; Ives, 1960; and Graul, 1960). It is equally apparent that the last two remaining ice sheets, Greenland and Antarctica, must have been in a condition of hydrologic equilibrium during this entire 6000 years. This

FIGURE 5.

(A) Eustatic curve of the last 20,000 years, based upon radiocarbon dates and submarine morphology from stable regions (Fairbridge, 1961). Oscillations over the last 6000 years are named for characteristic localities of raised beaches, drowned forests, and other things. For the earlier oscillations, names are chosen from the best-known stages of the glacial and periglacial sections of North America and Europe. Danish pollen zones are added for correlation.

(B) Sedimentation rate for total $CaCO_3$ from a tropical (3° S lat.) Atlantic deep-sea core by Wiseman and Todd (1959). Time is controlled by radiocarbon dating of end members and in detail from accurate analysis of Fe and Ti accumulations rates (believed constant). In the strictly equatorial belt, the $CaCO_3$ production probably reflects surface-water temperature (Emiliani, 1961), but in higher latitudes a different pelagic fauna and increased circulation during glacial periods may reverse this tendency.

confirms the opinion of certain glaciologists (Péwé, 1958; and others) and is not negated by *local* evidences of a certain lack of balance.

Calculations of the present volume of world continental icecaps vary from about 20 to 40 $\times$ 10^6 km.^{3} (Cailleux, 1954) but, following the recent seismic surveys of Antarctica, I have recalculated the figure at 37.5 $\times$ 10^6 km.^{3} (Fairbridge, 1961), although admittedly the order of accuracy is not high (± 5) owing to very large areas remaining unsurveyed. The melting inertia of such a body of ice due to latent heat is tremendous, especially in high latitudes where the summer albedo exceeds 90 per cent and there is a winter of total darkness. It is hardly surprising that it shows little reaction to short-period climatic cycles. Nye (1957) calculated at least a 5000-year delay from his study of ice-flow velocities.

Examination of the residue of the world's glacier ice discloses that it is almost entirely disposed in the form of mountain glaciers. Their geographic distribution is interesting. According to Flint (1957) their area is as follows:

Lat. 80 to 60° N:	Canadian Arctic	150,000 km.^{2}
	Alaska, Yukon	70,000 km.^{2}
	Northern Scandinavia	5000 km.^{2}
	Iceland	12,000 km.^{2}
	Spitsbergen, Norway	58,000 km.^{2}
	Novaya Zemlya in U.S.S.R.; and others	58,000 km.^{2}
		353,000 km.^{2}
Lat. 50 to 30° N:	Alps	3600 km.^{2}
	Caucasus	2000 km.^{2}
	The Himalayas, Tibet, and others	100,000 km.^{2}
		105,600 km.^{2}
Lat. 30 to 50° S:	Chile and Patagonia in South America	25,000 km.^{2}
	New Zealand	1000 km.^{2}
		26,000 km.^{2}
	Grand total	484,600 km.^{2}

It is immediately apparent that nearly 95 per cent of the mountain glaciers of the world lie in the Northern Hemisphere. Observations by glaciologists on selected units from well distributed parts of the world show that melting is gaining over accretion in the majority of them. The melt rate naturally varies greatly, but generally it is of the order of 1 to 10 $\times$ 10^5 m.^{3} km.$^{-2}$ years^{-1}, which is just of the right magnitude to produce a sea-level rise of about 1 mm. a year (0.36 $\times$ 10^{12} m.^{3}).

It is concluded therefore that, as Antarctica and Greenland are hydrologically almost static, variations in the rest of the world's glacier ice control the eustatic contemporary sea level. Short-period variations in the world's climate are therefore reflected by reactions in the mountain glacier economics. Furthermore, since 95 per cent of the world's mountain glaciers are located in the middle latitudes of the Northern Hemisphere, we conclude that these constitute the *sensitive latitudes* to climatic change (see FIGURE 7).

The fact that the Southern Hemisphere is so poorly represented in these latitudes means that the mechanical type of effective solar radiation (affected

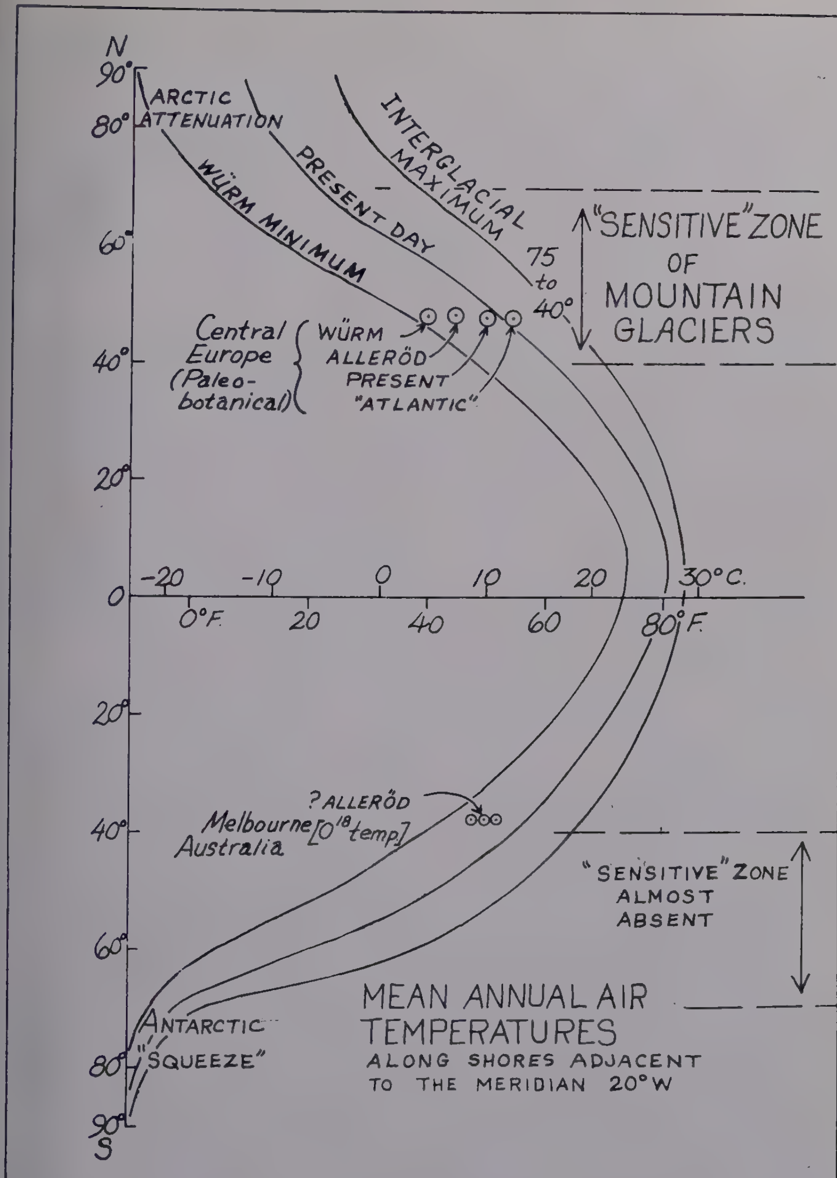


FIGURE 7. Diagram illustrating location of the "sensitive latitudes" of mountain glaciers in the Northern but not the Southern Hemisphere. Three curves indicate the present-day mean annual temperature gradient from the equator to the poles, and the approximate curves for glacial minima and interglacial maxima; these last are based upon available paleobotanical 0–18 isotope temperatures. Note that in the north the curves attenuate while, in the south, the presence of the semipermanent antarctic icecap causes them to converge. The steepening of the equator-pole thermal gradient in the Northern Hemisphere during cold phases is associated with a general acceleration of circulation systems, atmospheric and oceanic.

by the tilt of the earth and other such factors) is asymmetric in its meteorologic reactions. The southern sensitive latitudes will show the temperate effects of moderating oceanic currents, while the northern sensitive latitudes will react to the warmer radiation cycles with higher temperatures and to cooler ones with far-lower mean temperatures. The contrasts between high-index (strong zonal) and low-index (meridional) circulations will also be much more marked in the north.

A preliminary attempt (FIGURE 7) has been made to plot the hemispheric distribution of mean annual temperatures at sea level approximately about the meridian 20° W. Three curves are constructed: present day, interglacial maximum, and glacial (Würm/Wisconsin) minimum. Evidence is obtained chiefly from paleobotanical data (pollen, fossil leaves) and faunal ecologies. Noteworthy is the high degree of contrast and also attenuation in the Arctic, owing to the alternate freezing or melting of the oceanic regime (Brooks, 1949). In the Antarctic the opposite is true; the thermal curves are squeezed together at latitudes 60 to 70° S, owing to the steep mountainous barrier and the permanent freeze of the continental interior.

The fact that the Southern Hemisphere shows climatic changes that are almost entirely controlled by northern reactions is proved geologically by many data. Eustatic changes of sea level must be world-wide by definition. They are known to be high in the north because of warm interglacials, and they should also be associated in the south with warm faunal and floral indicators if the great climatic changes are synchronous; this is found to be true. A recent series of late Würm/Wisconsin and early Holocene oxygen isotope paleotemperatures determined at Melbourne, Australia, by Dorman and Gill (1959*a* and *b*) showed clearly the rising world temperature of the time (see FIGURE 7). The contemporary climatic change is world-wide (Kraus, 1960; Mitchell, 1961; and Schell, this monograph) and Dzerdzeevskii (1961) has demonstrated clearly that two quite distinct climatic epochs are discernible from the early 20th century data: a cooling one, followed by a marked warming one; a third cooling epoch began in the 1940 to 1950 decade, recognized as early in the United States records (Kincer, 1946) and now as world-wide (Wallén, personal communication).

In the Southern Hemispheres, the reactions are generally milder, but recent mountain glacier history of North and South America is approximately the same (Heusser, 1961) and the advances and retreats of glacier ice over the last century in New Zealand coincide broadly with those of Scandinavia (Kolb, 1953, 1958). Those of the late Pleistocene are demonstrated by C-14 dates (Brodie, 1957).

Precise coincidence of minor climatic events in both hemispheres is not expected because sunspot-induced pressure, precipitation, and temperature anomalies are not world-wide, nor necessarily synchronous (Willett, 1949, 1950; Fairbridge, 1961); thus it is normal to find certain glaciers advancing while others retreat (Heusser, 1956, and this monograph).

Croll-Milankovitch perturbation effects. As recognized by Croll nearly a century ago and as developed by Milankovitch into an elegant mathematical theory (for latest modifications, see van Woerkom, 1953; Zeuner, 1959), the astronomic perturbations of the earth's orbit and axis are measurable parameters that are susceptible to harmonic analysis. The effects would range

from varying the length of the seasons in alternate hemispheres to changing the amplitude of the seasons. For a smooth and symmetrical earth the effectiveness of such changes would be slight, as pointed out by Simpson (1940) and others, since the hemispheres would be reciprocally exposed. However it appears that Simpson failed to note the "sensitive" latitudes of the Northern Hemisphere, with the mean about 65° N, illustrated by Zeuner (1959), in which zone the effective solar radiation under astronomic perturbations was calculated to range about 4° C. above and below the present (slightly above the mean), having a mean frequency of about 40,000 years (see FIGURE 8.)

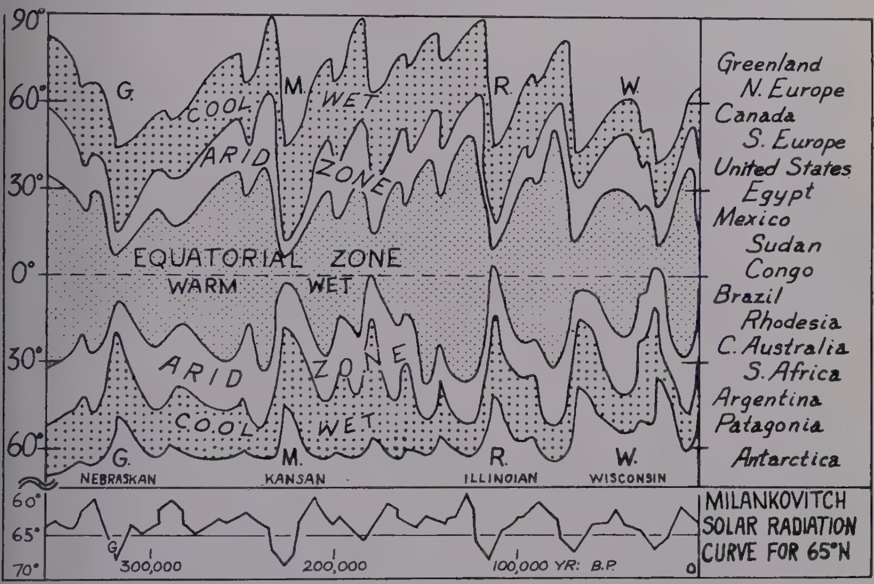


FIGURE 8. Reconstruction of the Quaternary world climatic zones, based upon the Emiliani-Milankovitch time scale; a similar treatment using the Soergel-Milankovitch system would be twice as long, but would have similar characteristics. Note the condensation of climatic zonal migration in the Southern Hemisphere, and its great extension in the Northern Hemisphere. Designed to illustrate the Penck hypothesis of expansion and contraction (due to reduced evaporation during glacials: Flohn, 1953) and rejecting the Büdel hypothesis of glacial pluvialism.

It has been said that the proof of synchronism of the Northern and Southern Hemisphere glaciations was fatal to the Croll theory, but this is not necessarily correct when we consider that the sensitive latitudes of the north essentially control the world pattern. It seems that the Milankovitch figures for 65° N, integrated with the other figures required by long-period sunspot changes, eustatic changes, and others, offer a solution of approximately the right order of magnitude to match the observed data.

In the last decade, further refinements have appeared. The Hungarian scientist Bacsák (1955) developed the concept of four basic "solar climatic types" that would develop from the interference, positive or negative, of the four main variables in Milankovitch's theorem. This stated that

$$I = \Delta W_s \Delta_\epsilon - m(e \sin \pi)$$

where Δ_ϵ and $\Delta(e \sin \pi)$ are variables of known period in celestial mechanics, which ΔW_s and m are variables related to geographic latitude, but not dependent upon time; in these ΔW_s increases toward the pole and m toward the equator. A cool summer phase is caused by Δ_ϵ and $\Delta(e \sin \pi)$ being in phase, which occurred 85 times in 600,000 years. In typical middle latitudes, the resultant climatic reactions are as follows: both are minima ($- -$), the result is glacial; with both maxima ($+ +$), the result is antiglacial; at the transition ($+ -$), the result is subtropical; and the fourth ($- +$), the result is subarctic. The term glacial leads (in middle latitudes) to cool summers and mild winters; antiglacial, to warm summers and hard winters; the other terms are self explanatory.

The Milankovitch theory created a new problem, namely that, prior to the Pleistocene, there was no continental glaciation, whereas a simple oscillation such as its author depicted should go on indefinitely. Bacsák's new calculations failed to disclose extended glacial phases in the Pliocene. Still earlier, there are the polar migrations and geotectonic changes that would also mitigate against such reactions. The revision, sometimes known as the Croll-Milankovitch-Bacsák theory, provides us with a calculated sequence of climatic types, back for 600,000 years. The correlation is partly based upon Soergel (1937) but, in any case, is at present merely a suggestion. This detailed analysis now requires world-wide testing, and is therefore reproduced here (TABLE 1).

Melt Retardation

Croll, Soergel, Zeuner and others have emphasized the role of retardation in maintaining a glacial condition, once established, by a series of short-period thermal "dips." The retardation referred to is the delay in ice melting, conditioned by the latent heat and other factors of a gigantic ice mass. The late Pleistocene is the one period where there are sufficiently precise C-14 datings and paleotemperature indicators to quantify the phenomenon.

One of the most interesting points of the late-Pleistocene/Holocene oscillation is that the rise of air and sea-water temperature was well in advance of the "Flandrian" sea level rise. The mean air temperature in western Europe had risen 3 to 4° C., from 16,000 to 12,000 B.P., and 7 to 8° C. (that is, 1 to 2° C. higher than today) by 8600 years B.P., while sea level was still respectively 30 and 15 m. below that of today.

It has been demonstrated beyond reasonable doubt that sea level reached its present stand ±3 m. by 6000 B.P. However, the mean annual air temperature in the midlatitudes (derived by fossil data) had already reached the present level by 10,000 B.P. There is thus a "eustatic retardation" of 4000 years.

As with all melting/freezing problems (phase changes), there is evidence of a critical threshold, dependent upon two principal variables: available calories/time and the area/volume ratio of the ice masses concerned. Each ice mass constitutes a semi-independent system. During a warming phase, melting will proceed generally from the small to the large, and from the equator toward the poles. The completion of melting of one mass does not mean that another area will immediately take over. Each has its own threshold. Thus a melting sequence will be marked as a random, steplike progression. Since at any one time certain areas will always be marked by increased precipitation, there will

TABLE 1

BACSÁK CHRONOLOGY OF CLIMATIC TYPES FOR THE LAST 600,000 YEARS

Beginning year	Solar climate type	Duration	Amplitude in canonic units
−5,700	Subtropical, ice-free	5,700	+136
−11,300	Antiglacial, ice-free	5,600	+428
−16,300	Antiglacial, ice-covered	5,000	
−17,500	Subarctic, ice-covered	1,200	+45
−26,900	Glacial W_3, ice-covered	9,400	−456
−39,700	Subarctic, ice-covered	12,800	46
−40,300	Glacial, ice-covered	600	0
−53,900	Subtropical, ice-covered	13,600	+107
−57,200	Antiglacial, ice-covered	3,300	0
−66,500	Subarctic, ice-covered	9,300	−109
−72,700	Glacial W_2, ice-covered	6,200	−546
−77,900	Glacial W_2, ice-free	5,200	
−80,800	Subarctic, ice-free	2,900	0
−82,800	Antiglacial, ice-free	2,000	+468
−88,200	Antiglacial, ice-covered	5,400	
−99,700	Subtropical, ice-covered	11,500	+187
−100,400	Antiglacial, ice-covered	700	0
−110,600	Subarctic, ice-covered	10,200	−187
−117,000	Glacial W_1, ice-covered	6,400	−644
−122,000	Glacial W_1, ice-free	5,000	
−122,600	Subarctic, ice-free	600	0
−133,800	Antiglacial, ice-free	11,200	+529
−140,500	Subtropical, ice-free	6,700	+190
−146,000	Glacial RW, ice-free	5,500	−234
−158,300	Subarctic, ice-free	12,300	−138
−159,800	Glacial, ice-free	1,500	0
−170,200	Subtropical, ice-free	10,400	+248
−179,200	Antiglacial, ice-covered	9,000	+528
−182,000	Subarctic, ice-covered	2,800	0
−192,800	Glacial R_2, ice-covered	10,800	−643
−201,900	Subarctic, ice-covered	9,100	−399
−203,800	Antiglacial, ice-covered	1,900	+170
−215,800	Subtropical, ice-covered	12,000	+344
−221,300	Antiglacial, ice-covered	5,500	+518
−226,500	Subarctic, ice-covered	5,200	−170
−232,000	Glacial R_1, ice-covered	5,500	−676
−237,500	Glacial R_1, ice-free	5,500	
−243,200	Subarctic, ice-free	5,700	−60
−249,200	Antiglacial, ice-free	6,000	+250
−261,600	Subtropical, ice-free	12,400	+200
−263,700	Glacial, ice-free	2,100	0
−276,200	Subarctic, ice-free	12,500	−150
−284,500	Glacial MR_4, ice-free	8,300	−395
−287,500	Subtropical, ice-free	3,000	0
−297,900	Antiglacial, ice-free	10,400	+550
−303,800	Subtropical, ice-free	5,900	+200
−308,000	Glacial, ice-free	4,200	−387
−318,300	Subarctic, ice-free	10,300	−139
−325,800	Glacial MR_3, ice-free	7,500	−339
−328,700	Subtropical, ice-free	2,900	0
−337,500	Antiglacial, ice-free	8,800	+413
−342,500	Subtropical, ice-free	5,000	0
−346,700	Glacial, ice-free	4,200	−50
−355,300	Subarctic, ice-free	8,600	−100
−362,600	Glacial $MR_2 = M_3$, ice-covered	7,300	−300
−364,700	Subtropical, ice-covered	2,100	0
−370,800	Antiglacial, ice-covered	6,100	+405
−374,700	Antiglacial, ice-covered	3,900	
−382,600	Subtropical, ice-covered	7,900	+200
−384,700	Glacial, ice-covered	2,100	−270

TABLE 1—*Continued*

Beginning year	Solar climate type	Duration	Amplitude in canonic units
−394,600	Subarctic, ice-covered	9,900	−270
−404,000	Glacial $MR_1 = M_3$, ice-covered	9,400	−331
−405,800	Subarctic, ice-covered	1,800	0
−411,300	Antiglacial, ice-covered	5,500	+214
−419,200	Subtropical, ice-covered	7,900	+200
−426,200	Antiglacial, ice-covered	7,000	+235
−428,500	Subarctic, ice-covered	2,300	0
−438,800	Glacial M_2, ice-covered	10,300	−529
−447,600	Subarctic, ice-covered	8,800	−207
−449,000	Antiglacial, ice-covered	1,400	0
−460,700	Subtropical, ice-covered	11,700	+216
−466,400	Antiglacial, ice-covered	5,700	+481
−470,300	Subarctic, ice-covered	3,900	−50
−476,700	Glacial M_1, ice-covered	6,400	−601
−480,700	Glacial M_1, ice-free	4,000	
−488,500	Subarctic, ice-free	7,800	−339
−491,800	Antiglacial, ice-free	3,300	+100
−502,900	Subtropical, ice-free	11,100	+145
−507,800	Antiglacial, ice-free	4,900	+150
−517,500	Subarctic, ice-free	9,700	−200
−528,400	Glacial GM, ice-free	10,900	−200
−531,000	Subtropical, ice-free	2,600	0
−538,000	Antiglacial, ice-free	7,000	+409
−543,000	Antiglacial, ice-covered	5,000	
−546,100	Subtropical, ice-covered	3,100	0
−548,100	Glacial G_2, ice-covered	2,000	−479
−553,700	Glacial G_2, ice-free	5,600	
−564,200	Subarctic, ice-free	10,500	−365
−567,400	Glacial, ice-free	3,200	−460
−574,900	Subtropical, ice-free	7,500	+498
−579,700	Antiglacial, ice-free	4,800	+715
−585,000	Antiglacial, ice-covered	5,300	
−587,500	Subtropical, ice-covered	2,500	0
−590,100	Glacial G_1, ice-covered	2,600	−550
−595,100	Glacial G_1, ice-free	5,000	
−600,000	Subarctic, ice-free	4,900	−200

be a tendency for small negative eustatic swings to interrupt each upward curve. Graul (1960) recently pointed out that the northern European continental ice melted approximately 3000 years before the North American.

Since each ice mass melts semiindependently, it is obvious that the thin, low-elevation, low-latitude ice masses must go first. However, it is not clear how much would be lost at the distal fronts and how much by ablation over the surface. The geological evidence shows that the distal retreats did not begin until several thousand years after the solar radiation minimum (of Milankovitch). It has also been suggested that the pan shape of the two great remnant ice masses, Antarctica and Greenland, has something to do with their longevity (Fairbridge, 1960). The short-lived North American and Scandinavian masses attenuated distally, while the high-latitude masses are boxed in by high mountainous walls, owing to the continued isostatic loading of more than one-half million years.

The curve of melting (eustatic rise) is, excluding minor oscillations, a sinusoidal one, beginning slowly, steeply rising in the middle, and tapering off at

the end. After the long retardation effects, melting gained by its own momentum. When the melting really begins it is like a dam bursting. Sea level rises at a mean rate of 50 mm. a year for centuries on end. The superimposed sunspot cycle will periodically retard, but, at other times will amplify this rate. The mean rate of water influx was 18×10^{12} metric tons annually for the period 10,800 to 9000 B.P. The bulk of this mud-laden water would have been poured into the North Atlantic Ocean and, at an early stage, the input into the Gulf of Mexico may have been at about 5×10^{12} metric tons annually above the present normal rate. Geomorphic evidence of the torrentlike streams escaping from the periglacial lakes are to be seen at many points around the southern margins of the icecaps.

The revolutionary effect of this great volumetric and temperature change was early noticed, due to the C-14 dating of Atlantic oceanic cores (Ericson *et al.*, 1956, 1961). The foraminiferal species display an approximately 500-mile northerly shift in their habitats, owing to the warming of the surface waters. The 0–18 determinations suggest a warming of 6 to 10° C. (Emiliani, 1954, 1955, 1961). A decrease in pelagic sedimentation rate occurred. The crescendo came about 11,000 B.P. (Broecker *et al.*, 1960). The beginning of the "great melting" is reflected by a color change in certain globigerinal ooze cores (gray-white to pinky white) somewhat earlier (ca. 15,000 B.P.; Broecker, personal communication).

According to the Ewing-Donn theory of ice ages (1956, 1958), the 11,000 B.P. event marked an abrupt change of state in the Arctic Ocean, which had been an almost stagnant, ice-free lake during the low sea-level period of the Wisconsin. Deep-sea cores from the Arctic indicated (Donn *et al.*, 1959) that most of the sediment consisted of organic ooze or clean quartz sands and boulders that must have been transported by icebergs, but that the upper 10 to 20 cm. of the section is poor in sand but rich in foraminifera which, these writers say, must represent an ice-covered Holocene. The effect of such a freezing over of the Arctic in early Recent time would cut off free circulation of cold arctic water with the result that the Atlantic Ocean would rapidly rise in temperature. According to the Ewing-Donn hypothesis, this 11,000 B.P. event was sudden and catastrophic.

The rapidity of the rise of sea level about this time is undoubted, but there is evidence of oscillations that complicate the picture. In the ocean there are well-developed terraces at many intermediate levels between the Wisconsin low (at least −100 m.) and the present, and the radiocarbon dates of shells from these terraces give evidence of a progressive and oscillating rise of sea level (Fairbridge, 1960, 1961). On land there are clear indications that the climatic amelioration after the ice maximum was marked by similar oscillations. Broecker and Orr (1958) found that the level of melt waters in the Great Salt Lake rose and fell by hundreds of meters over this critical period. The pollen analysts have shown that there were sharp reversals of air temperature that permitted a swing in some places from subarctic to subtropical plants, and back again, several times.

It is believed therefore that the Ewing-Donn deductions about the timing of the Arctic-melting-freezing phenomena are in error (see also Emiliani, 1961; Schwarzbach, 1961). It is recognized that the organic productivity of the

largely ice-covered Arctic today is currently several orders of magnitude below that of the North Atlantic. However, 10 to 20 cm. of foraminiferal sediment are found above the glacially rafted quartz sands; their average radiocarbon age is 9500 years B.P., which would mean that they coincide with the maximum air-temperature ("climatic optimum") phase of the postglacial.

Air temperature reached the present level approximately 10,000 years B.P. and, in the middle latitudes, rose to 2.5° C. (annual average) above the present, returning nearly to the present or even cooler mean levels about 4500 years B.P. Reinterpreted in this light, one may reasonably conclude that the present ice cover of the Arctic is a feature of the Subboreal and Sub-Atlantic stages (that is, the last 4500 years); also that the 10 to 20 cm. of foraminiferal sands were early Holocene (essentially Atlantic) and corresponded to an ice-free phase of high Arctic productivity. In the present century there has been an Arctic warming trend of only about 3 to 4° C. that has opened up the ice cover of large sectors to summer shipping (Schell, 1961). A continuation of this trend for several centuries would probably complete the process, although ice floes would undoubtedly develop extensively during the winter seasons when, in any case, the constant darkness would greatly inhibit organic productivity. Around 80° N lat. at 6000 years B.P., mean annual temperatures may have been 5 or 6° warmer than today. At any relatively warm stage, icebergs could ferry erratics. During most of the Wisconsin, in my opinion, the Arctic was frozen over, stagnating the circulation and permitting organic black muds to accumulate (as on the floor of a glacial lake in winter), while the quartz sands would be transported in during temporary warm phases that were long enough to permit unfreezing of the Arctic, but not long enough to provide an interglacial warm stage, the retardation delay for which might exceed 5, or even 10,000 years (see the next section).

A Test of the Milankovitch Theory

A fundamental weakness of the Milankovitch theory is that, although early doubts about hemispheric synchronism are now removed and no matter how sound the mathematical basis, the actual thermal effects are based on calculations into which may have crept certain false assumptions. First, three distinct curves are combined: the precession of the equinoxes (21,000-year cycle), the obliquity of the ecliptic (40,000-year cycle), and the eccentricity of the orbit (92,000-year cycle). The formulae relating the ratio of the three contain interpretive elements. Milankovitch (1930, 1941) made allowance in his calculations for the land areas of each hemisphere and, indeed, for every 10° latitude, so that gross errors of earlier solar-radiation calculations were eliminated.

Our Holocene paleotemperature curve contains a mechanism for a limited test of the Milankovitch calculation. If we take the present time as a point of zero departure, both with respect to mean air temperature and solar radiation, it is possible to go back to 6000 B.P. ("climatic optimum") when the mean middle latitude air temperature was 2.5° C. above the present, as indicated by paleobotanical material. At this point in time, Milankovitch calculated that the effective summer radiation was also only 2.5° C., and the

mean annual radiation was 0.5° C. above the present. Our curve shows that the climatic optimum was also a crescendo of certain shorter period cycles, the effective amplitude of which would be of the order of 2° C., which would satisfy the observed departure (see FIGURE 9).

If one applies the Milankovitch curve for 65° N to the whole late Quaternary air temperature-eustatic curve, it can be seen that the effective radiation maximum falls at 10,000 years B.P., or very close to the time of maximum rate of eustatic rise and maximum rate of oceanic temperature rise. The last radiation minimum was at 25,000 years B.P., with a mean insolation −1.5° C. below the present, and an effective summer departure of −4° C. Both rising curves reached the present mean level at 16,000 B.P.

At this time, air temperature was still far below the present (about −6° C.) and thus one would have to assume a "temperature retardation" (of 6000 years) as well as a "eustatic retardation" behind the temperature. Such a conclusion is necessary, even if the timing is not quite correct. With about 30×10^6 km.^{2} of ice in the Northern Hemisphere at the time, a very considerable delay would be involved as the world air temperature rose, while the eustatic rise would still lag somewhat farther behind. If the radiation theory is approximately correct, then the sea-level rise was retarded 10,000 years with respect to effective radiation.

An important factor in delaying the ice melting is the isostatic rebound. The dating of raised beaches in Scandinavia and correcting these findings against contemporary eustatic levels provide us with data as to the rate and timing of the uplift. Around 10,000 to 11,000 B.P., maximal crustal uplift was at least 10 cm. year $^{-1}$, flattening exponentially so that in peripheral areas the rates are down to the order of 1 mm. year^{-1} or less (Fairbridge, 1961). This interior uplift would favor small glacial readvances, both through increased slopes acting as a stimulant to "dead" ice and also through the greater orographic effect. Simultaneously with uplift, the outer marginal areas show evidence of a collapse of Daly's "marginal bulge" and of further subsidence.

The physical shape of the great ice-covered continental masses was also a significant delay mechanism. The North American and Scandinavian icecaps were essentially sheetlike, thinning marginally across the southern plains. In contrast, Antarctica and Greenland are like deep pies with elevated mountainous margins. This pan shape, in my opinion, has been an important factor in retarding the melting of those great icecaps. If the solar radiation curve of Milankovitch is in any way significant, we are now 10,000 years beyond the last peak, so that there seems little chance of Antarctic melting during the Holocene epoch.

In using the Milankovitch astronomic curve as a basis for correlating the Pleistocene, many authors have followed Soergel (1925), thus obtaining a "long Pleistocene" record essentially related to an interpretation of central European (Alpine) glaciation and loess. However alternative treatment based on oscillations in deep-sea sediment records gives a "short Pleistocene" (Emiliani, 1955; Rosholt *et al.*, 1961). Another "short" version, based on the Alaskan glaciation, is offered by Karlstrom (1961). This geologic paradox in no way detracts from the significance of the astronomic curve but introduces a very important chronologic problem that awaits solution (see FIGURE 10).

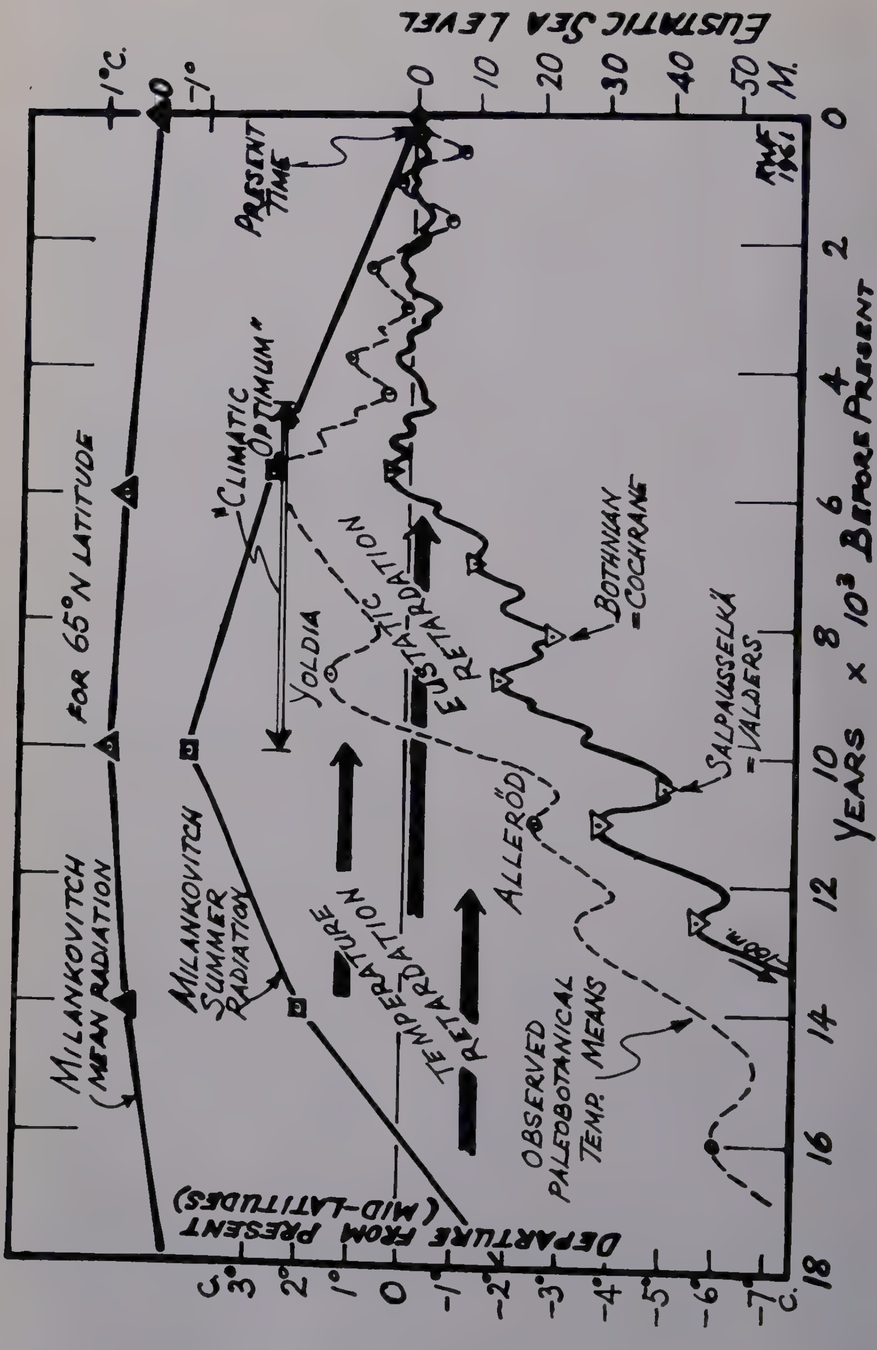

EUSTATIC SEA LEVEL
FOR 65°N LATITUDE
MILANKOVITCH MEAN RADIATION
MILANKOVITCH SUMMER RADIATION
PRESENT TIME
"CLIMATIC OPTIMUM"
YOLDIA
EUSTATIC RETARDATION
TEMPERATURE RETARDATION
ALLERÖD
BOTHNIAN =COCHRANE
SALPAUSSELKÄ =VALDERS
OBSERVED PALEOBOTANICAL TEMP. MEANS
DEPARTURE FROM PRESENT (MID-LATITUDES)
YEARS × 10³ BEFORE PRESENT

FIGURE 9.

An Eclectic Theory of Ice Ages

An eclectic theory of ice ages is now offered that embraces instrumentally demonstrated aspects of sea-level studies, meteorology, glaciology, and Quaternary geology:

(1) Polar wandering (or crustal shift) brought the South Pole into the mountainous continental mass of Antarctica, and the North Pole into the Arctic Ocean in late Tertiary times. I call this my "Polar Coincidence Theory." It is really only an amplification of Ramsay's (1924) "Relief Hypothesis."

(2) In middle Tertiary (early Miocene), the most important event in world oceanography was the closing of the Tethys Sea, a "Mediterranean" belt that extended from the East Indies and the Pacific through the present Mediterranean to the Atlantic. This essentially equatorial waterway was open for the last one-half billion years, and was now closed along an axis extending from Egypt to Anatolia, Turkey (the "Aleppo platform"). Soon afterward the Isthmus of Panama closed off the Atlantic from the Pacific. Cool oceanic currents now moved north and south in the Atlantic, replacing east-west warm currents; bottom temperatures dropped from about 8° C. to about 1° C. The Pacific remained warm until the Pleistocene (Emiliani, 1954), when gradually accelerated circulation brought successively colder waters into the tropical latitudes (Arrhenius, 1952, 1959).

(3) With the build-up of ice in Antarctica, world-wide lowering of sea level (and firm line) cooled the planetary atmospheric system by 0.5 to 1° C., the eustatic "control" of Zeuner (1959).

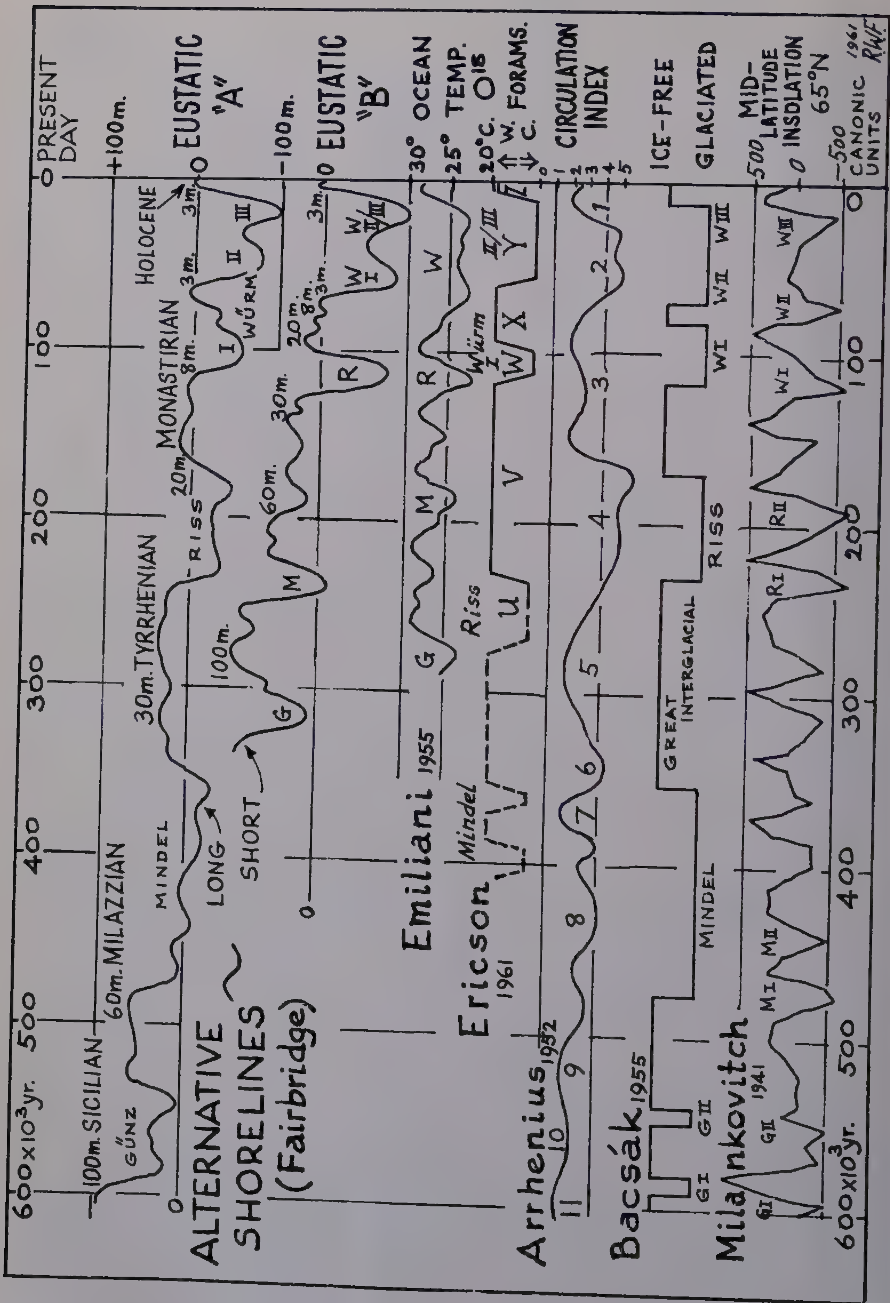
(4) Under astronomic solar radiation cycles in the sensitive 80 to 70° N latitudes, additional cooling by 1 to 1.5° C. cyclically depressed the firn line further, permitting certain mountain glaciers to merge into icecaps. This reinforcement of world trends began gradually with the beginning of the Pleistocene, each successive cool cycle carrying the system a stage further.

(5) Low intensity phases in the longer-period sunspot oscillations set up atmospheric circulation changes that permitted inclusion of certain mountain glacier centers into areas of higher precipitation and of lower temperatures (by 1 to 1.5° C.). Systematic circulation changes (of opposite trends) have been associated with the recent sunspot crescendo (Petterssen, 1949).

(6) Increased atmospheric disturbance, high-index circulation, resulting from the expansion of northern European and American icecaps and the consolidation of the Antarctic icecap, increased wind velocities in latitudes 10 to 40° N and S, and westerly storm tracks were diverted to more southerly courses

◄

FIGURE 9. The retardation principle applied to the record of the last 20,000 years. Four curves indicate: (A) eustatic changes of sea level, reflecting the total volume of ice melted, 1 mm. in sea level = 360×10^9 m.^{3} of melt water; (B) air temperature curve for typical middle-latitude region (northwestern and Central Europe), based upon the mean of large numbers of paleobotanical observations; (C) Milankovitch curve of effective radiation (insolation), for three summer months, thus the most effective ice-melting season; and (D) Milankovitch curve for mean annual radiation (note extremely small amplitude). Note that the eustatic curve is retarded with respect to air temperature, and the latter is delayed with respect to the insolation curve. Total retardation is approximately 10,000 years. Note coincidence of thermal maximum with Milankovitch's calculated summer-insolation temperature; since 6000 B.P. saw the last major ice remnant melted from North America and Scandinavia, since when the sea level has only changed in response to minor climatic cycles. However, Greenland and Antarctica have not melted and remain essentially static; their retardation factors (for high latitudes) probably exceed 50.000 years.



PRESENT DAY
600×10³yr.
EUSTATIC "A"
EUSTATIC "B"
HOLOCENE
MONASTIRIAN
TYRRHENIAN
MILAZZIAN
SICILIAN
GÜNZ
MINDEL
RISS
WÜRM
ALTERNATIVE SHORELINES (Fairbridge)
LONG
SHORT
30° OCEAN
25° TEMP.
20°C. O18
W. C. FORAMS.
Emiliani 1955
Ericson 1961
CIRCULATION INDEX
Arrhenius 1952
ICE-FREE
GLACIATED
GREAT INTERGLACIAL
Bacsák 1955
MID-LATITUDE INSOLATION 65°N
CANONIC UNITS
Milankovitch 1941
600×10³yr.
1961 RWF

FIGURE 10.

to curl up into Central Europe, Siberia, and the United States Midwest (Harmer, 1901; Eckardt, 1909; and Enquist, 1916). Evidence of such increased wind velocities is seen in the immense loess deposits of the midwestern United States, Eastern Europe, and China, and by eolian sands in Florida, the United States Gulf states, the Sahara, from South Africa to the Congo, and in Arabia, India, and Australia (see discussion by Jentzsch, 1908; Fairbridge and Teichert, 1953).

(7) Initially the Arctic basin must have been open, and marginal land masses would have been nourished by its moisture (as indicated by Ewing and Donn, 1956, 1958). Drop of sea level reduced the accessibility of warm currents, and freezing of the Arctic occurred. Tectonic uplift of the Wyville Thompson Ridge may have helped this condition (Rukhin, 1960). Ablation of the now very extensive ice sheets would be of very low order, in spite of greatly reduced nourishment, and there is no reason to suspect an immediate or rapid wasting of the North American or Eurasiatic ice sheets.

(8) With each interglacial epoch, under the incidence of Milankovitch radiation oscillations (effectively 40,000- to 90,000-year peaks), a 1 to 1.5° C. mean annual rise of temperature occurred in the "sensitive" middle latitudes, now the site of extensive although rather attenuated ice sheets. The effective summer radiation, however, rose by 4 to 5° C. above the mean. The longer-period sunspot effects (noted in paragraph 5) would augment the general zonal radiation in favored areas by 1 to 1.5° C. (mean annual rate). The development of anticyclonic centers over the southerly ice sheets would effectively reduce cloud cover and increase the insolation. The difference in the form of these Northern Hemisphere middle-latitude sheets and the pan-shaped Green-

◄

FIGURE 10. Seven curves to illustrate the dating of the Quaternary period. Note that the last 60,000 years for each are approximately identical, mostly being dated by radiocarbon. The extensions are by extrapolation, measured rates of sedimentation against thickness, and some newly developed uranium-isotope dating methods. However, the correlations with the standard glacial sequences are almost entirely hypothetical, and are based upon such coincidences as amplitude of oscillation and relative length, for example.

(A) Milankovitch, middle-latitude insolation. Correlation is based upon Soergel's understanding of the Penck and Brückner analysis of Alpine glaciation.

(B) Bacsák, climatic modification of insolation curve. Allows for retardation to maintain ice masses once developed, "bridging over" many smaller oscillations. Follows Soergel correlation.

(C) Arrhenius, circulation index, based upon production of $CaCO_3$ in Pacific deep-sea cores. High index (cold phase) is indicated by increased carbonate production (thus conversely from the equatorial Atlantic, probably owing to more open circulation: Arrhenius, 1959).

(D) Ericson (1961), foraminiferal ecology, in warm and cold zones (U, V, W, X, Y, Z) set against sediment thickness of an equatorial Atlantic deep-sea core. No adjustments are made for differing rates of sedimentation. Ericson (personal communication) notes that the relative thickness of interglacial layers drops about 50 per cent between the equator and 60° N lat.

(E) Emiliani, Atlantic ocean temperature, based upon 0–18 isotope determinations of pelagic foraminifera. Extrapolation from radiocarbon dating has recently been confirmed by ionium dating (Rosholt *et al.*, 1961). Correlation (hypothetical) with Alpine glaciations is suggestive of a "short Pleistocene."

(F) Fairbridge, eustatic "B" based upon Emiliani's 1955 curve, but modified to allow for lower interglacial sedimentation rates (Broecker *et al.*, 1958), and to show relative positions of interglacial high sea levels (Fairbridge, 1961).

(G) Fairbridge, eustatic "A", based upon Bacsák's 1955 calculations of climate that correspond to the Soergel "long Pleistocene" record. The graph also shows possible positions of interglacial "raised beaches." The correlations of curves F and G are designed merely as working hypotheses.

land and Antarctic sheets is very significant; the thin margins of the former melt much more easily than do the mountain-walled types (Fairbridge, 1960).

(9) Interglacial melting of the continental ice sheets proceeded initially from south to north, and there is no evidence that the Arctic Ocean was open until about 12,000 B.P., or that melting was influenced earlier by the presence of warm northerly waters. In the early Holocene melting, the smaller ice sheet of Scandinavia disappeared 3000 years before that of North America (Graul, 1960). Periodic readvances of the ice, during each retreat, reflect the secondary role of 550–1100–1650-year (sunspot?) cycles that led to short-term interruptions in the melting (Karlstrom, 1961).

(10) The rise of sea level came *pari passu* with the melting, but there must have been retardation delays, both in air temperature with respect to the effective radiation, and melting with respect to air temperature, in the order of several thousand years. I put the total eustatic retardation at 10,000 years (see FIGURE 9).

(11) The melting was extremely rapid, once set in motion. The southerly ice fronts were retreating at a mean rate of 2 to 3 m. per day during the summer seasons (Matschinski, 1960), leaving annual moraines (in central Sweden, for example) every 150 to 300 m. For several centuries at a time, for example around 10,000 B.P., the melt waters were pouring into the Atlantic Ocean at a phenomenal rate—18×10^{12} metric tons annually—raising the mean world level by 50 mm. a year. The marked lithologic change in all Atlantic deep-sea cores at or shortly before the beginning of the recent or Holocene epoch is thus a logical consequence of the melting, which proceeded most vigorously from south to north and drained predominantly into the North Atlantic.

(12) Greenland and Antarctica did not melt during the radiation oscillations of the last one-half million years, in part owing to their basin shapes (preventing gradual melting of attenuated distal margins available in the other ice sheets), and in part owing to the tremendous albedo of the high-latitude ice. In short, the retardation factor of a high-latitude icecap must exceed the length of the warm oscillations (hemicycles of less than 20,000 years), so that the next cool phase began before the ice sheet had warmed sufficiently to begin large-scale melting.

(13) Glacial climaxes and interglacial warm phases alternated through the Pleistocene, but not according to a strict or simple cycle. The Milankovitch-Köppen-Soergel or Emiliani curves for 65° N offer ready explanations, in so far as the three major astronomic cycles periodically fall out of phase. Thus, for example, the Hoxnian or Yarmouth interglacial shows a double high but a suppressed low, and has long been known as the "great interglacial" in view of its double length. The Würm or Wisconsin glacial, on the other hand, has at least two low peaks separated by a suppressed high (the Göttweig or Aurignacian interstadial).

(14) Prior to the Quarternary period, sunspot, and Milankovitch cycles were probably effective at certain times in producing well-known cyclic sedimentation sequences ("nonglacial varves," for example), but they did not fall into phase with a world-wide cooling phenomenon, except at a few well-marked stages: late Precambrian-early Cambrian (ca. 700 to 600 $\times 10^6$ years B.P.; both hemispheres), Siluro-Devonian (ca. 400 to 350; only in the Southern

Hemisphere), Permocarboniferous (ca. 280; only in the Southern Hemisphere), and Quaternary (both hemispheres). The occurrence of two of the above only in one hemisphere is explained by the paleomagnetic data that place the contemporary North Pole in the landless North Pacific.

(15) The fact that cooling and warming tendencies have been detected throughout geologic time seems to favor solar controls superimposed on the topographic mechanism as an explanation for climatic change. Actual "ice ages", however, called for the coincidence of topographic relief with high latitudes.

(16) Quantitative relating of sea-level changes to climatic change permits a close correlation between the hydrologic exchange and meteorologic controls (Fairbridge, 1960*a*, 1961; Wexler, 1961). Sea level may be regarded as the volume gauge of the world hydrologic balance.

(17) The Croll-Milankovitch-Bacsák correlation of major astronomic cycles (perturbations of the earth's attitude in the solar system) with climatic changes finds a large measure of support in the data from sea-level changes. These calculations deduce variations of the mean annual effective radiation in the sensitive 65° N lat. zone of ±1.5° from the norm. Such figures are of the same order as the observed 11-year (and longer) sunspot deviations, but may be superimposed at times on the latter to augment or nullify their local effect. If the Milankovitch curve is correct, the retardation factor in the warming of air temperatures in western Europe was about 4000–6000 years (due essentially to the albedo on the waning ice); the world rise of sea level was retarded further with respect to the air temperature by about 2000 to 4000 years (ice-melting delays were due to high volume/area ratios, the latent heat principle, and other such factors).

(18) Support is thus offered to some aspects of various ice-age theories that involve meteorological-topographic-eustatic controls (Croll, Milankovitch-Bacsák, Zeuner, Ewing and Donn, Emiliani and Geiss). The data thus demonstrated seem to provide an answer to certain meteorologists who still toy with the Simpson hypothesis of pure radiative control that totally neglects the vast body of available geological and glaciological evidence: to quote from John von Neumann (1960, p. 11), "The one important question that no one has been able to answer so far by dialectic methods is whether the ice age was due to the fact that the sun became hotter, or that the sun became cooler."

Summary

Quantitative paleoclimatology opens a new frontier in earth science, extending back more than 50,000 years. Time parameters are provided by radiocarbon dating, reappraisal of de Geer's varve chronology, by correlation with palynology, paleontological analysis, and by geochemical work on deep-sea cores. Climatic means (temperature, precipitation, and others) are provided by oxygen isotope studies, and by paleobotanical and pedologic indicators. Preliminary conclusions indicate that ice ages are produced by normal variations in observed solar-controlled meteorological effects that are reinforced, probably aperiodically in geologic time, by terrestrial, topographic "accidents."

Long-term meteorological and historical records disclose multiple cycles, for example 11, 22, 40, 80, 189, and 567 year periods, in part related to sunspots,

geomagnetic phenomena, and planetary motions. Mean annual temperatures in temperate belts vary 2 to 3° C.; the longer the period the greater the effect, in terms of snow-field accumulation or glacier ice melting. This concerns the world hydrologic balance and thus is reflected on world tide gauges. However, while accumulation advances directly in short steps, melting shows marked retardation.

Major astronomic cycles (in celestial mechanics), such as the precession of the equinoxes, with periods of 21,000, 40,000, and 92,000 years, control absolute and hemispheric variations in effective solar radiation. Although perhaps exaggerated by Milankovitch, middle-latitude temperatures will vary 3 to 5° C., but mainly in the Northern Hemisphere, which contains 95 per cent of world's sensitive mountain glaciers. Cyclicity of major climatic change is controlled thus by the Northern Hemisphere's topographic accidents (high mountains and broad continents).

Paleomagnetic data (Runcorn, 1956, 1959; and others) show secular polar migration through geologic history. Coincidence of poles with mountainous continents must initiate abnormal world-wide cooling, lowered snow lines, negative eustasy, and increased continentality. Such conditions occurred as rare accidents, only four times in the last 1×10^9 years, when they have been reinforced by mountain building, and further negative eustasy by oceanic trough deepening. Oceanic currents may be blocked, for example, as in Middle East and Panama during the middle-Tertiary; from then until early-Quaternary the middle-latitude temperatures dropped by 10 to 12° C.

The earth will remain relatively cool while Antarctica lies symmetrically about the South Pole. However, temperate belts will continue to be alternately glaciated and deglaciated every 40,000 to 90,000 years.

References

ABBOT, C. G. & F. E. FOWLE. 1913. Volcanoes and climates. Smithsonian Misc. Coll. **60** (29): 1–24.

ABBOT, C. G. 1957. Weather and solar variation. J. Solar Energy Sci. & Eng. **1:** 3–5.

ANDERSON, C. N. 1939. A representation of the sunspot cycle: Bell System Tech. J. **18:** 292–299.

ANDERSSON, G. 1902. Hasseln i Sverige, fordom och nu. Sverig. Geol. Undersokn. Stockhom, Afh. ser. ca. no. 3, 168 pp.

ANDERSSON, G. 1910. Swedish climate in the late-quaternary period. *In* Die Veränderungen des Klimas seitodem Maximum der letzten Eiszeit. Stockholm (11th Int. Geol. Congr.). : 247–294.

ÅNGSTRÖM, A. 1939. The change in the temperature climate in present time. Geogr. Ann. (Stockholm). **21:** 119–131.

ARRHENIUS, G. 1952. Sediment cores from the East Pacific. Rept. Swedish Deep Sea Expedition (Goteborg). **5**.

ARRHENIUS, G. 1959. Climatic records on the ocean floor. New York, Rockefeller Inst. Press (Rossby Memorial Volume). : 121–129.

BACSÁK, G. 1955. Pliozan- und Pleistozänzeitalter im Licht der Mimmelsmechanik. Acta Geol. (Acad. Sci. Hung., Budapest). **3:** 305–346.

BAKKER, J. P. 1958. Transgressionsphasen und Sturmflutfrequenz in den Niederländen in historischer Zeit. 21. Dt. Geographentag Wurzburg, Tagungsbär. u. Wiss. Abh. : 232–237. Wiesbaden, Germany.

BARGHOORN, E. S. 1953. Evidence of climatic change in the geologic record of plant life. *In* Climatic Change. H. Shapley, Ed. pp. 235–248. Harvard Univ. Press. Cambridge, Mass.

BARTA, G. 1956. A 40–50 Year Period in the Secular Variation of the Geomagnetic Field. Acta Geol. (Budapest). **4:** 15–42.

BAUR, F. 1948. Einführung in die Grosswetterkunde. Dietrichsche Verlag, 165 pp. Wiesbaden, Germany.

BENNEMA, J. 1954. Holocene movements of land and sea-level in the coastal area of the Netherlands. Geol. en Mijnb., no. 2. **16:** 254–262.
BERG, L. 1940. First Russian charts of the Caspian Sea. Bull. Acad. Sci. U.S.S.R. (Moscow), Geol. & Geophys. Ser. no. 2. : 159–180. (In Russian, with English summary.)
BERRY, E. W. 1922. A possible explanation of Upper Eocene climates. Proc. Am. Phil. Soc. **61:** 1–14.
BROECKER, W. S. & P. C. ORR. 1958. Radiocarbon chronology of Lake Lahontan and Lake Bonnerville. Bull. Geol. Soc. Amer. **69:** 1009–1032.
BROECKER, W. S., K. K. TUREKIAN & B. C. HEEZEN. 1958. The relation of deep sea sedimentation rates to variations in climate. Am. J. Sci. **256:** 503–517.
BROECKER, W. S., M. EWING & B. C. HEEZEN. 1960. Evidence for an abrupt change in climate close to 11,000 years ago. Am. J. Sci. **258:** 429–448.
BRODIE, J. W. 1957. Late Pleistocene beds, Wellington Peninsula. New Zealand J. Sci. Technol., Sect. B. **38:** 623–643.
BROOKS, C. E. P. 1949. Climate Through the Ages. London (Ernest Benn) and New York (McGraw-Hill). 2nd ed., 395 pp.
BROOKS, C. E. P. 1951. Geological and historical aspects of climate change. *In* Compendium of Meteorology. Am. Meteorol. Soc. : 1004–1018.
BÜDEL, J. 1951. Die Klimazonen des Eiszeitalters. Eiszeitalter u. Gegenwart. **1:** 15–26.
BÜDEL, J. 1959. The periglacial-morphologic effects of the Pleistocene climate over the entire world. Internat. Geol. Review. **1**(3): 1–16. (Transl. of German original in Erdkunde. **7:** 249–266, 1953.)
BUTZER, K. W. 1958. Russian climate and the hydrological budget of the Caspian Sea. Rev. Canadienne de Géogr. **12:** 129–139.
BUTZER, K. W. 1958. Quaternary stratigraphy and climate in the Near East. Bonner Geogr. Abh. **24:** 1–157.
CAILLEUX, A. 1954. Ampleur des régressions glacioeustatiques. Bull. Soc. Geol. France, ser. 6. **4:** 243–254.
CAREY, S. W. 1958. A tectonic approach to continental drift. *In* Continental Drift (Symposium, Univ. Tasmania). : 177–355.
CHAMBERLIN, T. C. 1899. An attempt to frame a working hypothesis of the cause of glacial periods on at atmospheric basis. J. Geol. **7:** 545–584; 667–685; 752–787.
CHANEY, R. W. 1940. Tertiary forests and continental history. Bull. Geol. Soc. Am. **51:** 469–488.
CLAYTON, H. H. 1920. Variations in solar radiation and the weather. Smithsonian Misc. Coll. **71** (3).
COLLINSON, D. W. & S. K. RUNCORN. 1960. Polar wandering and continental drift: evidence from paleomagnetic observations in the United States. Bull. Geol. Soc. America. **71:** 915–958.
CROHN, P. W. 1959. A contribution to the geology and glaciology of the western part of Australian Antarctic Territory. Bureau Min. Research Geol. Geophys. (Canberra, Australia). Bull. **52.**
DEWEY, E. R. 1958. The length of the sunspot cycle. J. Cyclic Research. **7:** 79–91.
DIAMOND, M. 1958. Precipitation trends in Greenland during the past thirty years. J. Glaciol. (Cambridge). **3:** 177–180.
DONN, W. L., M. EWING & R. J. MENZIES. 1959. Characteristics of the late Quaternary Arctic Ocean. Intern. Oceanogr. Congr. Am. Assoc. Advancement Sci. Preprints. : 19–20.
DORF, E. 1959. Climatic changes of the past and present. Contr. Mus. Paleont. Univ. Michigan. **13**(8): 181–210.
DORMAN, F. H. & E. D. GILL. 1959*a*. Oxygen isotope palaeotemperature measurements on Australian fossils. Proc. Roy. Soc. Victoria. **71:** 73–98.
DORMAN, F. H. & E. D. GILL. 1959*b*. Oxygen isotope paleotemperature determinations of Australian cainozoic fossils. Science. **130:** 1576.
DURHAM, J. W. 1959. Palaeoclimates. *In* Physics and Chemistry of the Earth. Pergamon Press. New York, London. **3:** 1–16.
ECKARDT, W. R. 1909. Das Klimaproblem der geologischen Vergangenheit und historischen Gegenwart, Braunschweig, 183 pp.
EGYED, L. 1956. The change of the Earth's dimensions determined from palaeogeographical data. Geofisica pura e appl. **33:** 42–48.
EGYED, L. 1961. The expanding earth. Trans. N. Y. Acad. Sci. Ser. 2. **23**(5).
EMILIANI, C. 1954. Temperatures of Pacific bottom waters and polar surficial waters during the Tertiary. Science. **119:** 853–855.
EMILIANI, C. 1955. Pleistocene temperatures. J. Geol. **63**(6): 538–578.
EMILIANI, C. & J. GEISS. 1959. On glaciations and their causes. Geol. Rundschau. **46** (for 1957). p. 576–601.

EMILIANI, C. 1961. Cenozoic climatic changes as indicated by the stratigraphy and chronology of deep-sea cores of Globigerina-ooze facies. N. Y. Acad. Sci. Ann. **95**(1).
ENQUIST, F. 1916. Der Einfluss des Windes auf die Verteilung der Gletscher. Bull. Geol. Inst. Upsala. **14:** 1–108.
ERICSON, D. B. 1961. Pleistocene climatic record in some deep-sea sediment cores. Ann. N.Y. Acad. Sci. **95**(1).
ERICSON, D. B., W. S. BROECKER, J. L. KULP & G. WOLLIN. 1956. Late-pleistocene climates and deep-sea sediments. Science. **124:** 285–289.
ERICSON, D. B., M. EWING, G. WOLLIN & B. C. HEEZEN. 1961. Atlantic deep-sea sediment cores. Bull. Geol. Soc. Am. **72:** 193–286.
EWING, M. & W. L. DONN. 1956. A theory of Ice Ages. Science. **123:** 1061–1066.
EWING, M. & W. L. DONN. 1958. A theory of Ice Ages—II. Science. **127:** 1157–1162.
FAIRBRIDGE, R. W. 1952. The geology of the Antarctic. *In* The Antarctic Today (N. Z. Antarct. Soc. Wellington, N.Z.). **:** 56–101.
FAIRBRIDGE, R. W. & C. TEICHERT. 1953. Soil horizons and marine bands in the coastal limestones of Western Australia. J. Proc. Roy. Soc. N.S.W. **86:** 68–87.
FAIRBRIDGE, R. W. 1955. Warm marine carbonate environments and dolomitization. Tulsa Geol. Soc. Digest. **23:** 39–48.
FAIRBRIDGE, R. W. 1959. Periodicity of eustatic oscillations (Periodizitaet der eustatischen Oszillationen). Int. Oceanographic Congr. (Washington, D.C., Am. Assoc. Advancement Sci.). **:** 97–99.
FAIRBRIDGE, R. W. 1960. The changing level of the sea. Sci. American. **202**(5)**:** 70–79.
FAIRBRIDGE, R. W. 1961. Eustatic changes in sea level. Physics and Chemistry of the Earth. **4:** 99-185. Pergamon Press. London, England.
FEDOROV, P. V. & L. A. SKIBA. 1960. Holocene levels of the Caspian and Black Sea basins. Izvestya Akad. Nauk, SSSR, ser. geograf. (July–Aug.), p. 24–34 (in Russian).
FLINT, R. F. 1957. Glacial and Pleistocene Geology. Wiley. New York, N.Y. 553 pp.
FLOHN, H. 1952. Allgemeine atmosphärische Zirkulation und Paläoklimatologie. Geol. Rundschau. **40:** 153–178.
FLOHN, H. 1953. Studien über die atmosphärische Zirkulation in der letzten Eiszeit. Erdkunde (Bonn). **7:** 266–275.
FLOHN, H. 1958. Bemerkungen zum Problem der globalen Klimaschwankungen. Arch. Meteorol. Geophys. u. Bioklimatol. Ser B. **9:** 1–13.
FRIEDMAN, H. 1961. Solar variability in X-ray and ultraviolet emissions observed by means of rockets. Ann. N.Y. Acad. Sci. **95**(1).
GEER, E. H. DE. 1959. La première datation pour un an précis d'un changement de climat. Cahiers Géol. **53:** 513–515.
GEIKIE, J. 1894. The Great Ice Age. 3rd ed. Stanford. London, England.
GILLULY, J. 1949. Distribution of mountain building in geologic time. Bull. Geol. Soc. Am. **60:** 561–591.
GLEISSBERG, W. 1958. The eighty-year sunspot cycle. J. Brit. Astron. Assoc. **68:** 148–152.
GRANLUND, E. 1932. De Svenska Högmossarnas Geologi. Sverig. Geol. Unders. Årsbok. **26**(1)**:** 1–193 (German summary).
GRAUL, H. 1960. Der Verlauf des glazialeustatischen Meeresspiegelansteiges, berechnet an Hand von C-14 Datierungen. Deut. Geographentag Berlin, Tagungsber. u. wiss. Abh. **:** 232–242.
HARMER, F. W. 1901. Influence of winds upon the climate of the Pleistocene. Quart. J. Geol. Soc. **57:** 405–478.
HAURWITZ, B. 1946. Relations between solar activity and the lower atmosphere. Trans. Am. Geophys. Union. **27:** 161–163.
HEEZEN, B. C. 1960. The rift in the ocean floor. Sci. American. **203**(4).
HENDERSON, E. P. 1959. A glacial study of central Quebec-Labrador. Geol. Survey Canada Bull. **50:** 94 pp.
HENNIG, A. 1911. Le conglomerat Pleistocene à Pecten. Wiss. Ergebn. Schwed. Sudpolar Exped., 1901–3. **3**(10).
HEUSSER, C. J. 1956. Postglacial environments in the Canadian Rocky Mountains. Ecol. Monogr. **26:** 263–302.
HEUSSER, C. J. 1961. Some comparisons between climatic changes in northwestern North America and southern Chile. Ann. N. Y. Acad. Sci. **95**(1).
HOBBS, W. H. 1911. Characteristics of Existing Glaciers. Macmillan. New York, N.Y.
HOYLE, F. 1955. Frontiers of Astronomy. Heinemann. London, England.
HUMPHREYS, W. J. 1913. Volcanic dust as a factor in the production of climatic changes. J. Wash. Acad. Sci. **3:** 365–371.
HUNTINGTON, E. 1907. The historic fluctuations of the Caspian Sea. Bull. Am. Geogr. Soc. **39:** 577–596.

IRVING, E. 1956. Paleomagnetic and paleoclimatological aspects of polar wandering. Geofis. pura e appl. **33:** 22–41.

IVES, J. D. 1960. Permafrost in central Labrador-Ungava. J. Glaciol. **3:** 789–790.

JENTZSCH, A. 1908. Über den Eiswind und das Dünengebiet zwischen Warthe und Netze. Monatsber. deut. geol. Ges. **:** 120–123.

KARLSTROM, T. 1961. The glacial history of Alaska: its bearing on paleoclimatic theory. Ann. N.Y. Acad. Sci. **95**(1).

KATS, N. Y. 1960. The climate of the Pleistocene epoch in relation to the development of ice sheets. Izvestiya Vsesoyuznogo Geograficheskogo Obshchestva, Jan./Feb. (in Russian).

KINCER, J. B. 1946. Our changing climate. Trans. Am. Geophys. Union. **27:** 342–347.

KLUTE, F. 1930. Verschiebung der Klimagebiete der Letzten Eiszeit. Petermanns Mitt. Erg. Heft. **209**.

KLUTE, F. 1935. Allgemeine Länderkunde von Afrika. Hannover, Hahnsche Buchhandlung (Allgem. Länderkunde d. Erdteile, pt. 3), 298 pp.

KOLB, A. 1953. Fluctuations of glaciers and climate on New Zealand during the last hundred years. Eighth Pac. Sci. Congr. (Philippines, 1953), Abstracts pp. 44–45.

KOLB, A. 1958. Historische Gletscherschwankungen auf der Südhalbkugel insbesondere auf Neuzeeland. Schlenschriften. **190:** 123–146.

KÖPPEN, W. 1914. Lufttemperatur, Sonnenflecken und Vulkanausbrüche. Meteorol. Zeitschr. **31**.

KÖPPEN, W. & A. WEGENER. 1924. Die Klimate der geologischen Vorzeit. Berlin (Gebr. Bortraeger), 256 p.

KRAUS, E. B. 1955. Secular changes of tropical rainfall regimes. Quart. J. Roy. Meteorol. Soc. **81:** 198–210.

KRAUS, E. B. 1956. Secular changes of the standing circulation. Quart. J. Roy. Meteorol. Soc. **82:** 289–300.

KRAUS, E. B. 1960. Synoptic and dynamic aspects of climatic change. Quart. J. Roy. Meteorol. Soc. **86:** 1–15.

KUIPER, G. P. 1957. The Planet Earth. Chap. 2. D. R. Bates, Ed. Pergamon. London, England.

LABRIJN, A. 1945. Het klimaat van Nederland gedurende de laatste twee en een halve eeuw. Kon. Nederl. Met. Inst., Meded. Verh. (s'Gravenhage). **49**(102)**:** 1–114. (W. Engl. summary.)

LAMB, H. H. & A. I. JOHNSON. 1959. Climatic variation and observed changes in the general circulation. Geogr. Ann. (Stockholm). **41:** 94–134.

LOEWE, F. 1956. Contributions to the glaciology of the Antarctic. J. Glaciol. **2**(19)**:** 657–665.

LVOVICH, M. I. 1961. The water balance of the land: Soviet Geography. **2**(2)**:** 14–27 (Transl. From Russian of 3rd Congr. Geogr. Soc. USSR).

MANLEY, G. 1953. The mean temperature of central England, 1698–1952. Quart. J. Roy. Meteorol. Soc. London. **79:** 242–261.

MATSCHINSKI, M. 1960. La "lenteur" des phénomènes glaciaires. Compt. rend. somm. Soc. Géol. France (1960), **:** 30–31.

MELLOR, M. 1959. Mass balance studies in Antarctica. J. Glaciol. **3:** 522–533.

MILANKOVICH, M. 1930. Mathematische Klimalehre und Astronomische Theorie der Klimaschwankungen. Handbuch Klimatolotie, Berlin. **1A**, 176 pp.

MILANKOVICH, M. 1941. Kanon der Erdbestrahlung und seine Andwendung auf das Eiszeitenproblem. Acad. Roy. Serbe (Belgrade), ed. sp. **133**. Sect. Sci. Math. Natl. 33.

MITCHELL, J. M., JR. 1961. Recent secular changes of global temperature. Ann. N.Y. Acad. Sci. **95**(1).

NAIRN, A. E. M. 1960. Paleomagnetic results from Europe. J. Geol. **68:** 285–306.

NEUMANN, J. VON. 1960. Some remarks on the problem of forecasting climatic fluctuations. *In* Dynamics of Climate. R. L. Pfeffer, Ed. **:** 9–11. Pergamon Press. Oxford, England.

NYE, J. F. 1957. The distribution of stress and velocity of glaciers and ice sheets. Proc. Roy. Soc. (London) Ser. A. **239:** 113–133.

ÖPIK, E. J. 1953. On the causes of the Paleoclimatic variations and of the ice ages in particular. J. Glaciol. **2:** 213–218.

PEARSON, H. W. 1901. Oscillations in the sea-level. Geol. Mag. Dec. IV. **8:** 167–174, 223–231, 253–265.

PETTERSSON, O. 1914. Climatic variations in historic and prehistoric time. Svenska Hydrogr.-Biol. Komm., Skriften. **5**.

PETTERSSEN, S. 1949. Changes in the general circulation associated with the Recent climatic variation. Geogr. Ann. (Stockholm). **31**. Glaciers and Climate. **:** 212–221.

PÉWÉ, T. L. 1958. Quaternary Glaciation: McMurdo Sound Region. Natl. Acad. Sci. IGY Bull. No. **14**.

PLASS, G. N. 1959. Carbon dioxide and climate. Sci. American. **201:** 41–47.
PLASS, G. N. 1961. The influence of infrared absorptive molecules on the climate. Ann. N. Y. Acad. Sci. **95**(1).
POST, L. VON. 1946. The prospect for pollen analysis in the study of the earth's climatic history. New Phytologist. **45:** 193–217.
RAMSAY, W. 1924. The probable solution of the climate problem in geology. Geol. Mag. **61:** 152–163.
RINGWOOD, A. E. 1960. Some aspects of the thermal evolution of the earth. Geochim. et Cosmochim Acta. **20:** 241–259.
RINGWOOD, A. E. 1961. Changes in solar luminosity and some possible terrestrial consequences. Geochim. et Cosmochim. Acta. **21:** 295–296.
ROSHOLT, J. N., C. EMILIANI, J. GEISS, F. F. KOCZY & P. WANGERSKY. 1961. Absolute dating of deep-sea cores by the Pa^{231}/Th^{230} method. J. Geol. **69**.
RUKHIN, L. B. 1960. Problem of the origin of continental glaciation. Intern. Geol. Rev. **2:** 925–935 (trans. from Russian of Izvestiya Vses. Geogr. Obsh., 1958).
RUNCORN, S. K. 1956. Magnetization of rocks. *In* Handbuch der Physik (Berlin, Springer Verlag). **47** pt. **1:** 470–497.
RUNCORN, S. K. 1959. Rock magnetism. Science. **129:** 1002–1012.
SCHOFIELD, J. C. 1960. Sea level fluctuations during the last 4000 years as recorded by a chenier plain, Firth of Thames, New Zealand. N. Z. J. Geol. Geophys. **3:** 467–485.
SCHOVE, D. J. 1955. The sunspot cycle, 649 B.C. to A.D. 2000. J. Geophys. Research. **60:** 127–146.
SCHWARZBACH, M. 1953. Orogenesen und Eiszeiten. Zur Ursache des Klimawechsels in der Erdgeschichte. Die Naturwissenschaften (Berlin). **17:** 452–455.
SCHWARZBACH, M. 1961. Das Klima der Vorzeit. F. Enke, 2nd ed., 275 p., Stuttgart, Germany.
SCHWARZSCHILD, M. 1958. Structure and Evolution of the Stars. Princeton Univ. Press. Princeton, N. J.
SIMPSON, G. C. 1940. Possible causes of change in climate and their limitations. Proc. Linn. Soc. London. **152:** 190–219.
SIMPSON, G. C. 1957. Further studies in world climate. Quart. J. Roy. Meteorol. Soc. **83:** (358).
SOERGEL, W. 1925. Die Gliederung und absolute Zeitrechnung des Eiszeitalters. Fortschr. Geol. u. Pal. **13:** 251 pp.
SOERGEL, W. 1937. Die Vereisungskurve. Berlin, 87 p.
STACEY, C. N. 1961. Cyclical measures—some tidal aspects concerning equinoctal years. Unpublished manuscript.
STILLE, H. 1924. Grundfragen der vergleichenden Tektonik. (Borntraeger). Berlin, Germany, 443 pp.
STUIVER, M. 1961. Variations in radiocarbon concentration and sunspot activity. J. Geophys. Research. **66:** 273–276.
TEICHMÜLLER, R. 1958. Die Niederrheinische Braunkohlenformation. Fortschr. Geol. Rheinland u. Westf. **2:** 721–750.
UMBGROVE, J. H. F. 1947. The Pulse of the Earth. Nijhoff, 2nd ed. The Hague, Netherlands. 358 pp.
UREY, H. 1952. The Planets. Yale Univ. Press. New Haven, Conn.
VISHER, S. S. 1921. Increased oceanic salinity as one cause of increased climatic contrasts. Bull. Geol. Soc. Am. **32:** 429–435.
WAGER, L. W. 1933. The form and age of the Greenland ice cap. Geol. Mag. **70:** 145–156.
WEIZSÄCKER, C. F. VON. 1943. Ueber die Entstehung des Planetensystems. Z. Astrophys. **22:** 319–355.
WEXLER, H. 1952. Volcanoes and world climate. Sci. American. **195:** (April).
WEXLER, H. 1961. Ice budgets for Antarctica and changes in sea-level. J. Glaciol. **3:** 867–872.
WILLETT, H. C. 1949. Solar variability as a factor in the fluctuations of climate during geological time. Geogr. Ann. (Stockholm), **31**. Glaciers and Climate. : 295–315.
WILLETT, H. C. 1950. Temperature trends of the past century. Roy. Meteorological Soc., Centenary Proc. : 195–206.
WILLETT, H. C. 1951. Extrapolation of sunspot-climate relationships. J. Meteorol. **8**(1): 1–6.
WILLIS, E. H., H. TAUBER & K. O. MÜNNICH. 1960. Variations in the atmospheric radiocarbon concentration over the past 1300 years. Am. J. Sci. (Radiocarbon Suppl. 2). : 1–4.
WISEMAN, J. D. H. & I. TODD. 1959. Signification des variations du taux d'accumulation de Globorotalia menardii (d'Orbigny) dans une carotte de l'Atlantique equatorial. Colloq. Intern. Cent. Natl. Rech. Sci. **83:** 193–208.

WOERKOM, A. J. J. VAN. 1953. The astronomical theory of climate changes. *In* Climatic Change. H. Shapley, Ed. Harvard Univ. Press. : 147–157. Cambridge, Mass.

WOLDSTEDT, P. 1954–1958. Das Eiszeitalter. 2nd ed. Erike Verlag. Stuttgart, Germany. 2 vols.

ZAKLINSKAYA, Y. D. 1958. The principles of paleobotanical methods of zonation of the Cenozoic deposits of Kazakhstan and adjacent parts of the West Siberian Lowland (in Russian). Izvestiya Akademii Nauk SSSR, ser. geol. **10:** 72–85.

ZEUNER, F. E. 1953. Das Problem der Pluvialzeiten. Geol. Rundschau. **41:** 242–253.

ZEUNER, F. E. 1959. The Pleistocene Period: Its Climate, Chronology and Faunal Successions. Hutchinson Sci. and Techn. London, England.

Part VII. Palynology, Dendrochronology, and Varve Chronology

SOME ASPECTS OF THE VARIANCE SPECTRA OF TREE RINGS AND VARVES*

Reid A. Bryson and John A. Dutton
Department of Meteorology, University of Wisconsin, Madison, Wis.

The Problem

Centuries-long climatic records, as gathered with standard meteorological instruments, are rather unusual except in Europe. This situation can be corrected only with a great deal of patient waiting and, while this will lengthen the records, it will not extend them back in time. The student of paleoclimatology will always be faced with the necessity of seeking substitute time series that are in some understandable way equivalent to the missing time series of climatic measurements. For decades, tree-ring thickness series have been used as such a substitute time series.

The nature of a substitute series must be carefully considered before conclusions based on it can be accepted, however. Two major problems may be posed: first, that of determining to which climatic series the tree-ring series is equivalent; and, second, that of interpreting the behavior of the substitute series. It is toward certain aspects of this second question that the following paragraphs will be largely directed. Specifically, we shall direct our attention to the cyclic behavior of tree-ring, varve, and sunspot series, and shall largely pass over the secular variation of these series. Both of these problems have been treated in the literature, but it appears that certain more recently developed statistical treatments might yield deeper insight and greater precision than those that have been used. The variance spectrum technique (power spectrum) is one of these powerful new tools for studying the cyclic behavior of series.

Method of Analysis

The theoretical basis. Variance spectrum analysis provides a statistically sophisticated method of determining the cyclical behavior of a time series, and avoids most of the difficulties present in older, more subjective techniques for finding hidden periodicities. Several significant parameters can be derived from a variance spectrum, and confidence intervals for the results can be calculated.

As will be mentioned later, the spectra increase in statistical significance as the amount of data increases, and on this basis the technique is well suited to long tree-ring records. A brief explanation of the mathematical theory follows.

Most scientists are familiar with the Fourier analysis of a periodic function, and a departure to variance spectra can readily be made from the Fourier

* This research was supported in part by the Office of Naval Research, Geography Branch, Washington, D.C.; the University of Arizona Institute for Atmospheric Physics, Tucson, Ariz.; the University of Wisconsin Numerical Analysis Laboratory, Madison, Wis.; and the Midwestern Universities Research Association, Madison, Wis.

theory. Consider a function $f(x)$ periodic in the interval $(0, 2\pi)$, and without loss of generality let $f(x)$ have zero mean. We can write the Fourier series for $f(x)$ (using a tilde to indicate the relation between a function and its series, but not implying approximation),

$$f(x) \sim \sum_{n=1}^{\infty} (a_n \cos n x + b_n \sin n x), \tag{1}$$

where a_n and b_n are given by well-known relations (Zygmund, 1935). Since f is of zero mean, we may write its variance and the variance of its series as

$$\frac{1}{2\pi}\int_0^{2\pi} f^2(x)\, d x \sim \frac{1}{2}\sum_{n=1}^{\infty} (a_n^2 + b_n^2)$$

using the orthogonality of trigonometric functions. If f is continuous, we have an exact relation known as Parseval's theorem (Rudin, 1953),

$$\frac{1}{2\pi}\int_0^{2\pi} f^2(x)\, d x = \frac{1}{2}\sum_{n=1}^{\infty} (a_n^2 + b_n^2). \tag{2}$$

The term $(a_n^2 + b_n^2)/2$ shows how much of the total variance of $f(x)$ is contained in sinusoidal waves of the nth harmonic; or, in other words, how much variance is contained in cycles of length $2\pi/n$. Clearly a sine function of period $2\pi/n$ will have all of its variance given by b_n^2/n, which indicates the importance of peaks in spectra. On the other hand, it can be shown that a time series composed of a collection of small random impulses of random duration ("white noise") has a constant spectral density (Lanning and Battin, 1956).

Decomposition of the variance of a time series into variance associated with sinusoidal waves of various frequencies is precisely what we wish to do in this paper, so that the periodicities in our specific series are discovered.

The generalization of these obviously valuable concepts to infinite, nonperiodic functions was accomplished by Wiener (1930), using the Fourier integral (or Fourier transform), which can be obtained formally by extending a Fourier series to an infinite interval (Wiener, 1932).

The following development of the theory is a heuristic one, but is based on the rigorous derivation. In particular, Wiener worked with an integrated spectrum, but his results reduce to those given here if we are interested only in the absolutely continuous component of the spectrum.

Wiener started with the assumption that the autocorrelation function

$$\varphi_{ff}(\tau) = \lim_{T\to\infty} \frac{1}{2T}\int_{-T}^{T} f(t)\, f(t+\tau)\, dt \tag{3}$$

exists, and demonstrated that the spectral density $\Phi_{ff}(\omega)$ (or spectrum, as the exact term is abbreviated in common usage) is given by the Fourier transform of φ_{ff},

$$\Phi_{ff}(\omega) = \frac{1}{\pi}\int_{-\infty}^{\infty} \varphi_{ff}(\tau)\, e^{-i\omega\tau}\, d\tau. \tag{4}$$

From Equation (3) it is clear that $\varphi(-\tau) = \varphi(\tau)$, so discarding the odd part of the integrand, which integrates to zero, we obtain

$$\Phi_{ff}(\omega) = \frac{2}{\pi}\int_0^\infty \varphi_{ff}(\tau)\cos\omega\tau\, d\tau. \tag{5}$$

Thus,

$$\Phi_{ff}(-\omega) = \Phi_{ff}(\omega).$$

Via the invertibility of Fourier transforms it is also true that

$$\varphi_{ff}(\tau) = \frac{1}{2}\int_{-\infty}^\infty \Phi_{ff}(\omega)\, e^{i\omega\tau}\, d\omega \tag{6}$$

and

$$\varphi_{ff}(\tau) = \int_0^\infty \Phi_{ff}(\omega)\cos\omega\tau\, d\omega. \tag{7}$$

Equation **7** can be used to demonstrate what information the spectrum gives. Since $\varphi(0)$ is the variance of f, and

$$\varphi_{ff}(0) = \int_0^\infty \Phi_{ff}(\omega)\, d\omega, \tag{8}$$

it is clear that $\Phi_{ff}(\omega)$ gives the portion of the variance contained in the frequency interval $(\omega, \omega + d\omega)$.

Formally combining Equations **3** and **4**, we obtain

$$\begin{aligned}
\Phi_{ff}(\omega) &= \lim_{T\to\infty}\frac{1}{2\pi T}\int_{-T}^{T}\int_{-T}^{T} f(t)\, f(t+\tau)\, e^{-i\omega\tau}\, dt d\tau \\
&= \lim_{T\to\infty}\frac{1}{2\pi T}\int_{-T}^{T}\int_{-T}^{T} f(t)\, e^{i\omega t} f(t+\tau)\, e^{-i\omega(t+\tau)}\, dt d\tau \\
&= \lim_{T\to\infty}\frac{1}{2\pi T}\int_{-T}^{T} f(t)\, e^{i\omega t}\, dt \int_{-T}^{T} f(t')\, e^{-i\omega t'}\, dt' \\
&= \lim_{T\to\infty}\frac{1}{2\pi T}\left|\int_{-T}^{T} f(t)\, e^{-i\omega t}\, dt\right|^2.
\end{aligned}$$

This is the analogue of Equation 2.

Before we proceed to the practical aspects of variance spectrum analysis it seems worthwhile to discuss briefly some of the assumptions involved. Wiener's development used an assumption of stationarity of the time series: in particular, the assumption that the autocorrelation depends only on the length of the lag interval, not the particular location in the series. This assumption is inherent in requiring the existence of an autocorrelation function dependent only on τ, as in Equation **3**. However, requiring the existence of the autocorrelation as a valid representation of a statistical property of the time series also implies that the ergodic property is valid because a time average and a probability average have already been interchanged.

If the stationary time series generated by a random or stochastic process has a Gaussian joint probability distribution with zero mean, the autocorrelation function (as a probability average) completely specifies the distribution; hence, so does the spectrum. If the process is stationary but not Gaussian, we still obtain much useful information from the autocorrelation, and the spectrum is a valid representation of the average variance associated with specific frequencies. Furthermore, although the results on the statistical reliability of spectra are exact for Gaussian processes, they are probably good approximations otherwise (Blackman and Tukey, 1958).

We can still compute a spectrum of a nonstationary time series, but its interpretation becomes a much more difficult task.

The greatest problem we face, however, is that our data is not a mathematical function or a function of time generated by a random process, but a time series with values at discrete points over a finite interval. We thus know really very little about the process we hope to study. It cannot be overemphasized that the spectra must therefore be interpreted, within the limits of statistical reliability, as valid only over the interval used in the computation. We may assume that the spectrum approaches the theoretical spectrum more closely as the length of the interval increases. In fact Kahn (1957) points out that this may be considered a theorem.

Computation and statistical reliability of variance spectra. The advantage of spectra derived from the autocorrelation method over the Fourier series technique is that we obtain estimates of average variance over a frequency interval rather than lines at discrete frequencies. Actual computation of a spectrum requires more care than a simple reduction of integrals to sums. The subject is covered in detail by Blackman and Tukey (1958), and the computation formulas used on the digital computer for this study are based on their work.

Since the computation formulas are expressed in terms of lags, as used in any autocorrelation study of discrete data, we must calculate the frequency or period in real time associated with a particular lag. Letting Δt be the interval between pieces of data in the original time series, the real period P for the j^{th} lag is given by

$$P = \frac{2m}{j}\,\Delta t \left[\frac{\text{years}}{\text{cycle}}\right] 0 \leq j \leq m$$

the frequency $\omega/2\pi$ is the reciprocal, $1/P$.

It is obvious that the longest period is the infinite one, and the shortest is $2\Delta t$, in accord with sampling theory. A spectral estimate at lag j is an average over the interval

$$\left(\frac{j}{2m\,\Delta t} - \frac{1}{2\Delta t}, \qquad \frac{j}{2m\,\Delta t} + \frac{1}{2\Delta t}\right)$$

although spectra are frequently plotted for convenience as straight-line segments from point to point rather than in histogram form.

Blackman and Tukey also discuss in detail the calculation of confidence limits for spectra. The method is based on a Chi-square distribution and the Gaussian assumption. For a large number N of pieces equally spaced data

and a small number m of total lags we may say that the number of degrees of freedom k is given by

$$k \cong \frac{2N}{m}.$$

As an example, for $k = 10$ from tables in Blackman and Tukey we can find that given a computed spectral estimate of value X at a relatively "smooth" place in the spectrum there is an 80 per cent probability that the true long-run estimate Y is in an interval

$$X/1.6 \leq Y \leq X/0.49.$$

Usually more important is the size differential between humps and valleys. For $k = 10$, we have 80 per cent confidence that humps are significantly different than valleys if the value at the top of the hump is greater than the value in the valley by a factor of 1.6/0.49. At sharp peaks we have only 2 degrees of freedom, and a peak-to-valley ratio of 2.3/0.10 for 80 per cent significance.

There exist several techniques for improving the statistical reliability of results, or for achieving the same reliability with less computation (Blackman and Tukey, 1958). The most useful of these, in view of the 2 degrees of freedom at peaks, is to remove the peaks by filtering the data (prewhitening), compute the spectra of the filtered data and, then, operate on the spectra by the appropriate function (variance transfer function) to put the peaks back into the spectrum. The necessary function can be computed from the filter used.

Such methods were not used on any of the data in this paper, as it was desired at this stage only to investigate the possibilities of spectral analysis of tree-ring series. The usefulness of this analysis seems well demonstrated by following sections of the paper. The spectra shown here can be used for designing prewhitening filters for the tree-ring series used and continuing the study.

If the variance spectra are plotted with spectral estimates as linear ordinates against frequency as linear abscissas, the areas under the curve are preserved so that we can compare various parts of the spectrum and, as demonstrated by Equation **8**, the total area is the variance of the series. Thus it is wise to plot against frequency as a linear co-ordinate and then convert the scale to the reciprocal periods.

Because of the wide range of frequencies we may wish to plot the estimates against the logarithm (to the base 10) of frequency. Note that

$$\int_{\omega_1}^{\omega_2} \Phi(\omega)\, d\omega = (\log_e 10) \int_{\log \omega_1}^{\log \omega_2} \omega\Phi' (\log \omega)\, d \log \omega$$

where it is understood that $\Phi' (\log \omega_j)$ is the same number as $\Phi(\omega_j)$. Thus, areas are preserved relative to each other if we plot $\omega\Phi(\omega)$ linearly against $\log_{10} \omega$. Introducing the factor $\log_e 10$ preserves the areas with respect to the original linear–linear plot.

Some of the graphs in this paper include a plot of $\log \Phi(\omega)$ against frequency, which aptly demonstrates changes of several orders of magnitude without scale changes.

A useful relation may be derived from log–log plots of spectra, although the relative amplitude of peaks is disguised. Many log–log plots show an approximately linear trend over much of the frequency range. Recalling that spectral estimates of a variable with units U have units of

$$\left[\frac{\text{variance}\,(U)}{\omega}\right] = \left[\frac{U^2}{\omega}\right]$$

we may derive a useful power law if the spectrum plots as an approximation of a straight line. Assuming for future reference that the slope of this line is $-5/3$, we may write

$$\frac{U^2}{\omega} \propto \omega^{-5/3}$$

or

$$U \propto \omega^{-1/3} = P^{1/3}.$$

Cross spectral analysis. The technique of variance spectral theory can be extended to study relationships between two different series. Again, proceeding formally, the cross correlation function of two functions f and g is

$$\varphi_{fg}(\tau) = \lim_{T\to\infty} \frac{1}{2T} \int_{-T}^{T} f(t)g(t+\tau)\, dt.$$

The cross spectrum is again a Fourier transform,

$$\Gamma_{fg}(\omega) = \frac{1}{\pi} \int_{-\infty}^{\infty} \varphi_{fg}(\tau)\, e^{-i\omega\tau}\, d\tau,$$

or

$$\Gamma_{fg}(\omega) = \Phi_{fg}(\omega) + i\,\Psi_{fg}(\omega) = \frac{1}{\pi} \int_{-\infty}^{\infty} \varphi_{fg}(\tau)(\cos \omega\tau - i \sin \omega\tau)\, d\tau.$$

The real part, Φ_{fg}, is called the cospectrum; and the imaginary part, Ψ_{fg}, the quadrature spectrum. The cospectrum is a measure of the in-phase covariance between the two series, and the quadrature spectrum is a measure of the out-of-phase covariance between the two series. It is obvious that the phase angle between the two series is given by

$$\Theta(\omega) = \text{arc tan}\,[\Psi_{fg}(\omega)/\Phi_{fg}(\omega)].$$

What we wish to know is how well the two series are related at various periods. The question is answered by the coherence, R, defined by

$$R^2(\omega) = \Phi_{fg}^2(\omega) + \Psi_{fg}^2(\omega)/\Phi_{ff}(\omega) + \Phi_{gg}(\omega).$$

It is clear that the coherence equals 1 for $f = g$ ($\Psi_{ff}(\omega) = 0$), and is greater than or equal to zero. There is a method of finding the probability of the coherence exceeding a value R^* when it is really zero. Defining the degrees of freedom, $k_R = [(2N - m/2)/N]$, we have

$$R^* = (1 - p^{2/(k_2-1)})^{1/2}$$

where p is the probability level in question. As an example for 10 degrees of freedom there is only a 1 per cent chance that the coherence will equal or exceed 0.89 (Panofsky and Brier, 1958, following N. R. Goodman) if it is really zero. Thus we say a coherence value of 0.89 is significant at the 99 per cent level.

In summary then, we shall utilize variance spectra to study tree-ring series for the following reasons.

(1) This technique provides measures of the average variance in consecutive frequency bands; hence all periodicities, from the minimum allowed by sampling theory to an infinite one, are considered: not only those chosen a priori, as in the usual Fourier Analysis.

(2) Significant periodicities, if present, will appear as peaks on the spectral plot, and the significance of these peaks can be tested.

(3) The presence of power-law relations between variance and frequency may be readily discerned.

(4) The similarity of two series may be rigorously examined by use of the cross-spectrum, and the frequencies at which there is significant parallelism between series may be identified.

Results of Analysis

The tree-ring data used in this study came from one source: the University of Arizona dendrochronology group. It is therefore the same collection of data on which many previous studies using different methods have been based.

Two types of tree-ring data were used. Several long series of actual ring thicknesses from individual trees were examined, in part to explore longer periodicities. Most spectra calculated were for indices based on groups of trees (Schulman, 1956). There is a normal decrease in the annual radial growth increment, and considerable variance of a long-period nature is absent in the indices since this growth-rate decrease has been removed. Furthermore the nature of the fitting process by which the normal curve is removed tends to remove certain periods from the record, although it would require a separate inquiry to determine the precise statistical effect of this somewhat subjective process. In addition the well-known decrease of the variance of annual growth as the tree ages means that the spectra of young (200 to 1000 year) trees will exhibit more average variance per century than those of trees 2 or 3 milleniums in age.

As an example, the variance spectrum of actual annual ring thickness for a 3-millenium California sequoia is shown in FIGURE 1, and for comparison the spectrum of a 1500-year-old limber pine from the Snake River basin is shown in FIGURE 2. The most obvious feature of these two spectra is that while there are peaks and valleys at the higher frequencies (shorter periods), none stands out as dominant. To be sure, there are some peaks higher than others, but there are so many nearly the same that one can say only that for periods less than about a decade the variance is evenly distributed over many frequencies. The spectrum of FIGURE 2 illustrates the less uniform distribution of variance characteristic of the somewhat shorter series. Again, for periods less than a decade, the variance is spread through many frequencies, but a few

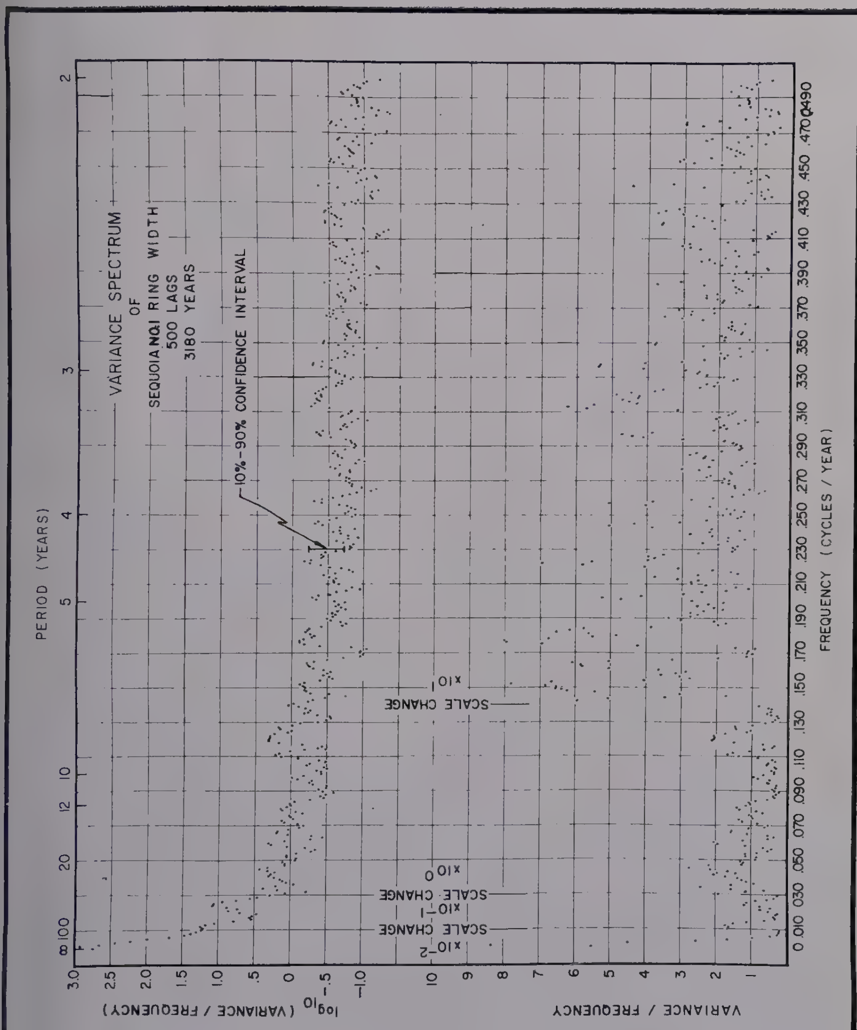


FIGURE 1. Variance spectrum of a 3-millenium California sequoia.

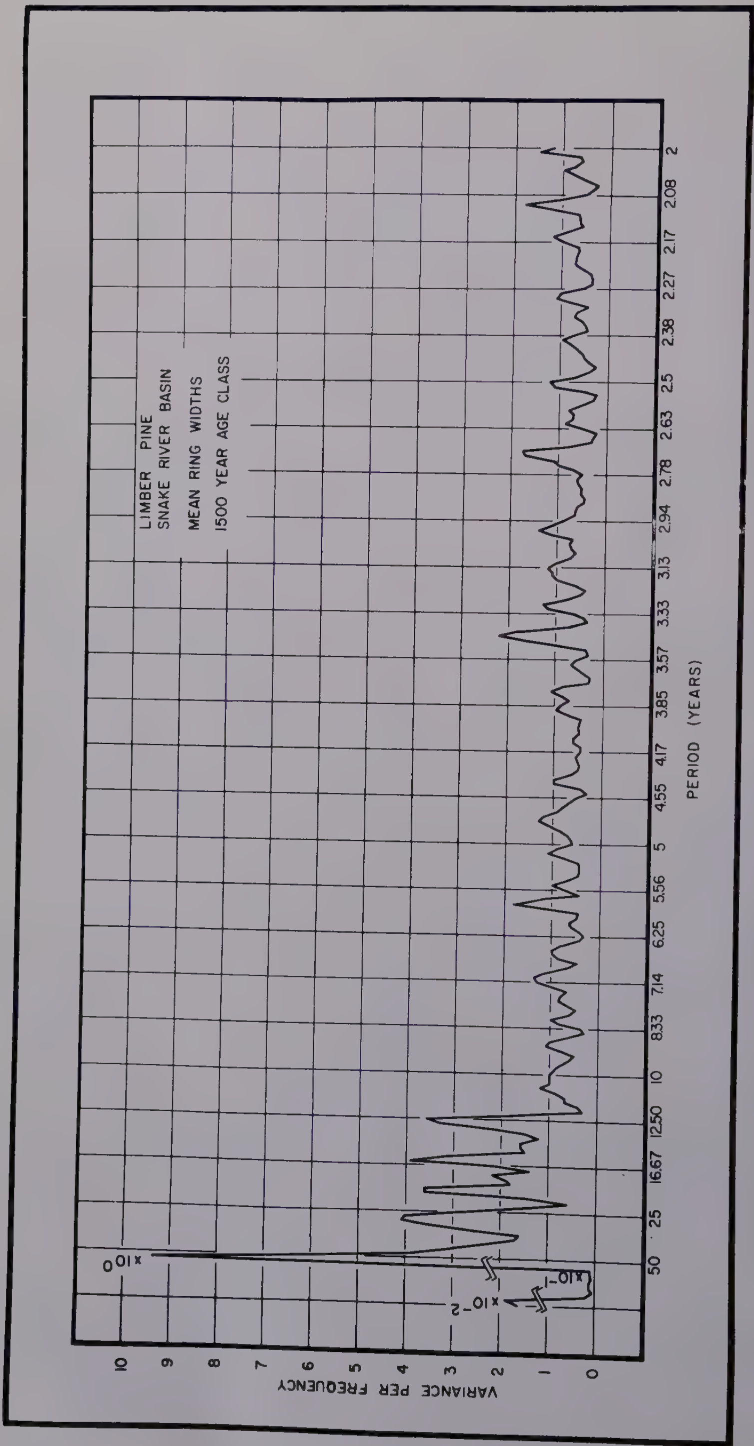


FIGURE 2.* Variance spectrum of a long-lived limber pine from the Snake River basin.

* Spectral estimates have been connected by straight line segments in this and other figures in this paper for convenience only. Histogram treatment is more appropriate.

stand out slightly. They are not significant at the 90 per cent level, however. These are at about 2.1, 2.7, 3.4, and 5.8 years, and are emphasized here because they tend to recur in many of the spectra to follow. At longer periods, peaks near 13, 16, 20, and 30 years appear, but it must be remembered that the time interval increases for these longer periods and the precision with which the particular period can be specified decreases. There is a pronounced change in FIGURE 2 between the generally decreasing variance as the period decreases from "infinity" to about 10 years, after which the variance is nearly inde-

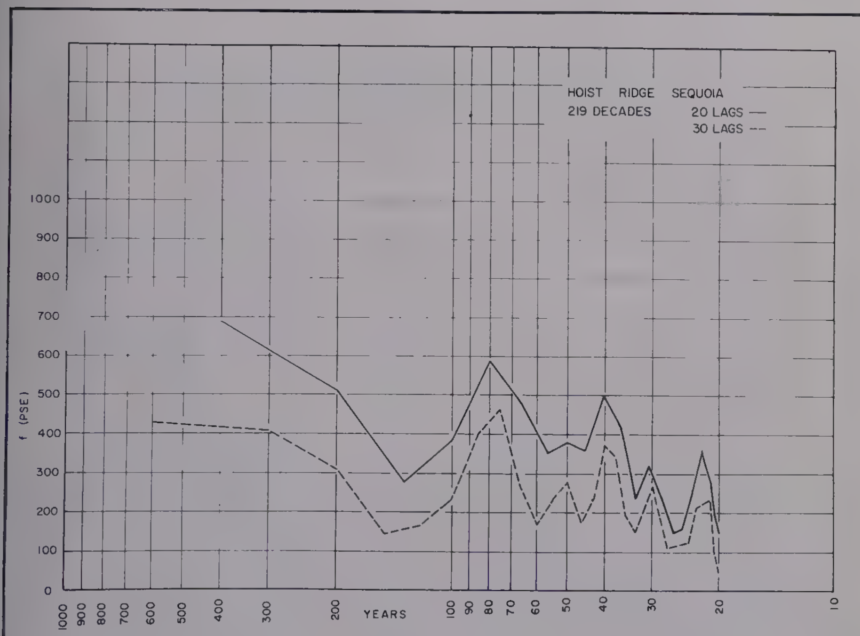


FIGURE 3. The variance spectrum of a 2190 ring series from a California sequoia. The 2 curves shown are for different averaging intervals. Since each point represents the variance within a frequency interval, the greater the number of such intervals, the smaller the variance. Thus, the 30-lag curve is below that for 20 lags. The label f (PSE) on the ordinate indicates that the spectral density estimate has been multiplied by the frequency for ease in seeing the spectral details.

pendent of period (frequency). This lack of dependence between variance per frequency interval and frequency suggests "white" noise.

The decade-average ring-thickness spectrum for a 2-millenium sequoia (FIGURE 3) exhibits long-period peaks similar to those found in the spectrum of the Snake River limber pine of FIGURE 2. The peaks of this particular spectrum from California are most pronounced near 23 years, 40 years, and 80 years: they are clearly not coincident with the Snake River basin peaks.

FIGURES 4 to 11 present the variance spectra of series of tree-ring indices based on groups of trees from various areas of western North America. The basic data for these spectra is from the work of Schulman (1956). There are certain similarities in all these spectra and the many others computed but not

presented here. There are also some regional similarities, and some regional differences. FIGURES 4 and 5, particularly, show a broad hump of variance for periods one to three decades in length in the northwestern United States. This hump diminishes northward, as shown on the spectrum of the Fraser River trees in Canada (FIGURE 6), and disappears to the east (Banff area, Alta., Canada, FIGURE 7) and to the south.

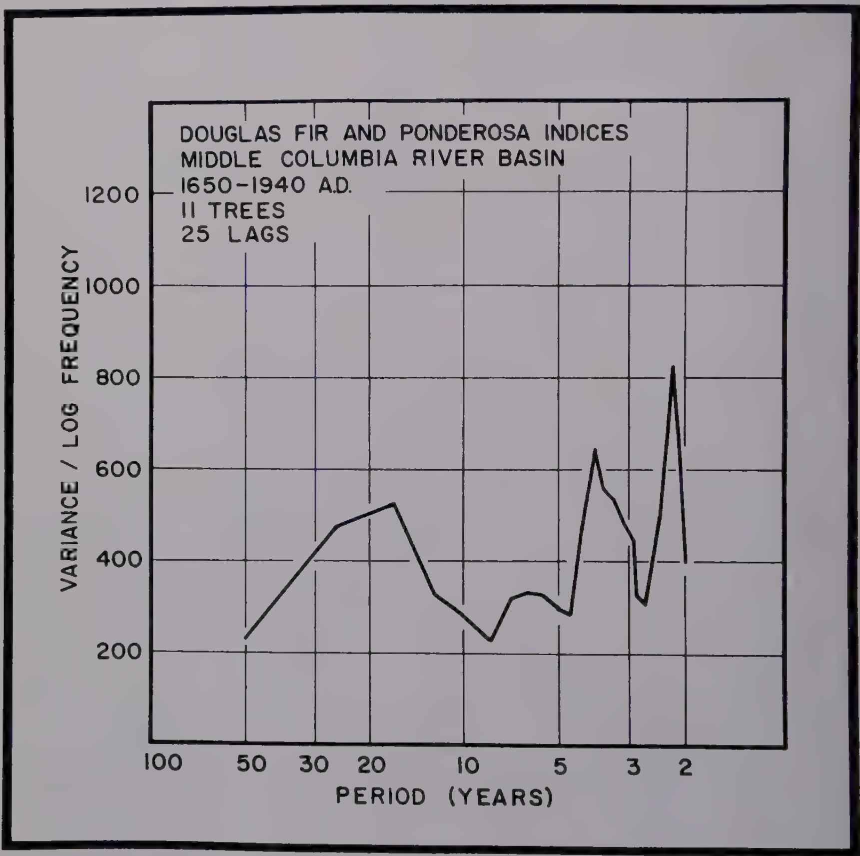


FIGURE 4. The variance spectrum of Douglas fir and ponderosa indices from the middle Columbia River basin.

The longest prominent periodicity in the North Platte and Colorado River basins appears to be from 4 to 6 years (FIGURES 8 and 9), but this hump is not important either to the north or south. Peaks at a little over 2 and a little over 3 years are the only characteristic of the Arkansas River basin spectrum (FIGURE 10). These same peaks are found in the Gila River (Arizona) basin spectrum, but with peaks at 5 to 6 and 8 to 9 years also present (FIGURE 11). The reader has probably noticed that the peaks in FIGURE 10 also appear in FIGURES 5, 7, and 9. In fact, one or the other of these peaks appears on nearly

every tree-ring spectrum computed. Both peaks were present on the spectrum of a set of Miocene tree rings.

For comparison with this North American data, the spectrum of a Patagonian (South America) ring index series was computed (FIGURE 12). The broad hump at 20 to 30 years is present, as in FIGURES 4 and 5, as are the peaks at 2+ and 3+ years and a hump at 6 to 7 years. One could readily class this

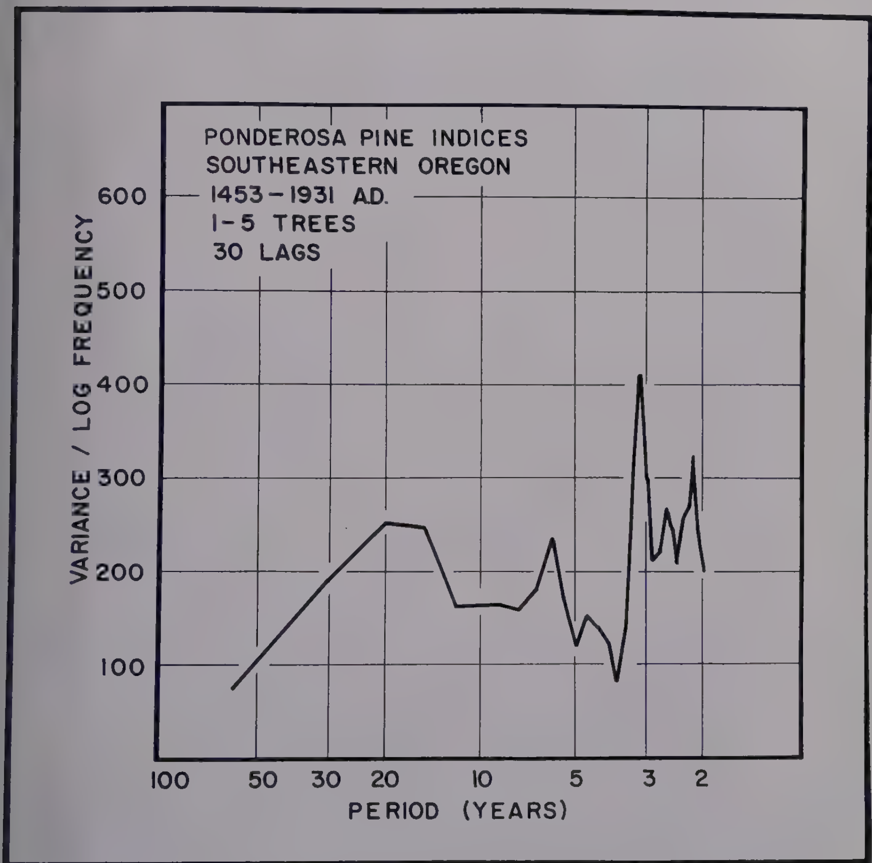


FIGURE 5. The variance spectrum of ponderosa indices from southeastern Oregon.

spectrum with those from the states of Washington and Oregon and, interestingly, the data are from similar climatic regions.

Three general features stand out regarding the spectra presented above. First, there is the relatively universal presence of the 2+- and 3+-year peaks, indicating somewhat more than random amount of variance associated with these periods. Although high levels of significance are only occasionally found for these peaks, their recurrence on many spectra leads one to suspect some reality to their existence. If these periodicities are of more importance than

others, their universal nature suggests an external, globally present cause (Wasserfall, 1930). Second, there are the occasionally significant humps and peaks that vary in period from area to area. These, if "real," cannot be due to causes entirely external to the area involved, although they may represent

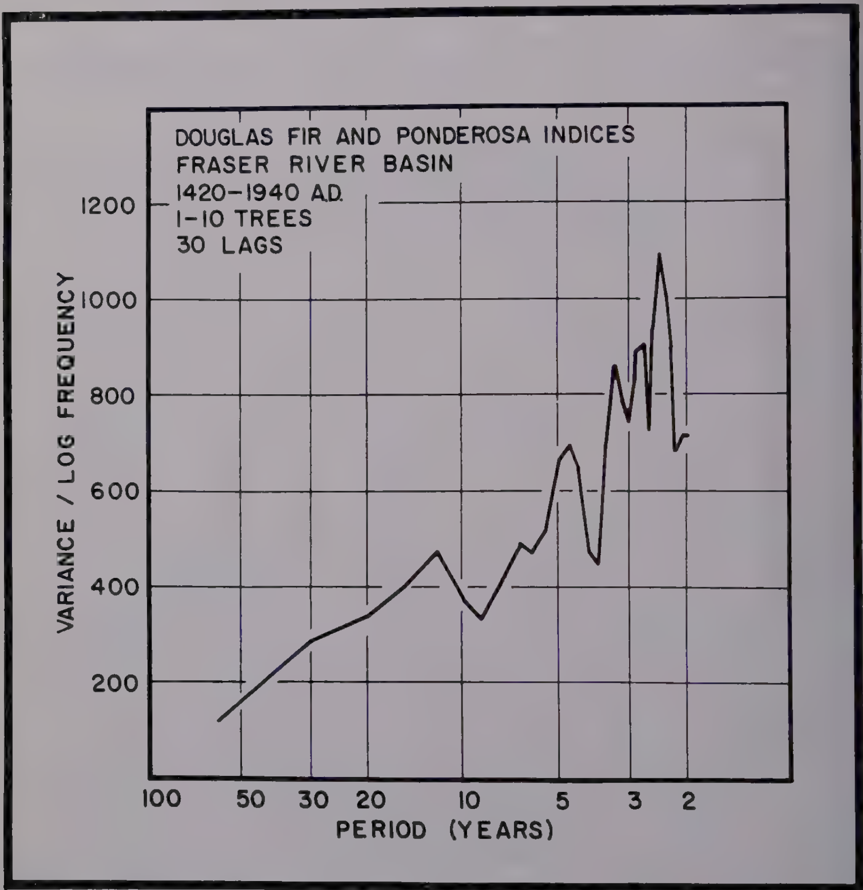


FIGURE 6. The variance spectrum of Douglas fir and ponderosa indices from the Fraser River basin.

the locally possible modes of response to a perhaps random external forcing function. Whether this is so, or whether these periodicities are of local origin, they must be regarded as essentially ecological in nature. They may represent a swinging of the balance between the trees and their competitors or enemies. If so, once the tree is 1000 years old it should be well enough established to dominate its surroundings, and these intermediate periodicities should diminish. Comparison of the spectra of a sequoia for its first and second millenium separately indicates that they actually do diminish.

Third, it will be noted that there is no tendency for a peak of variance at 11 years, and only a few spectra show humps or peaks near 22 years. In fact most tree-ring spectra exhibit a minimum of variance in the frequency band where sunspots show a maximum of variance (FIGURE 13).

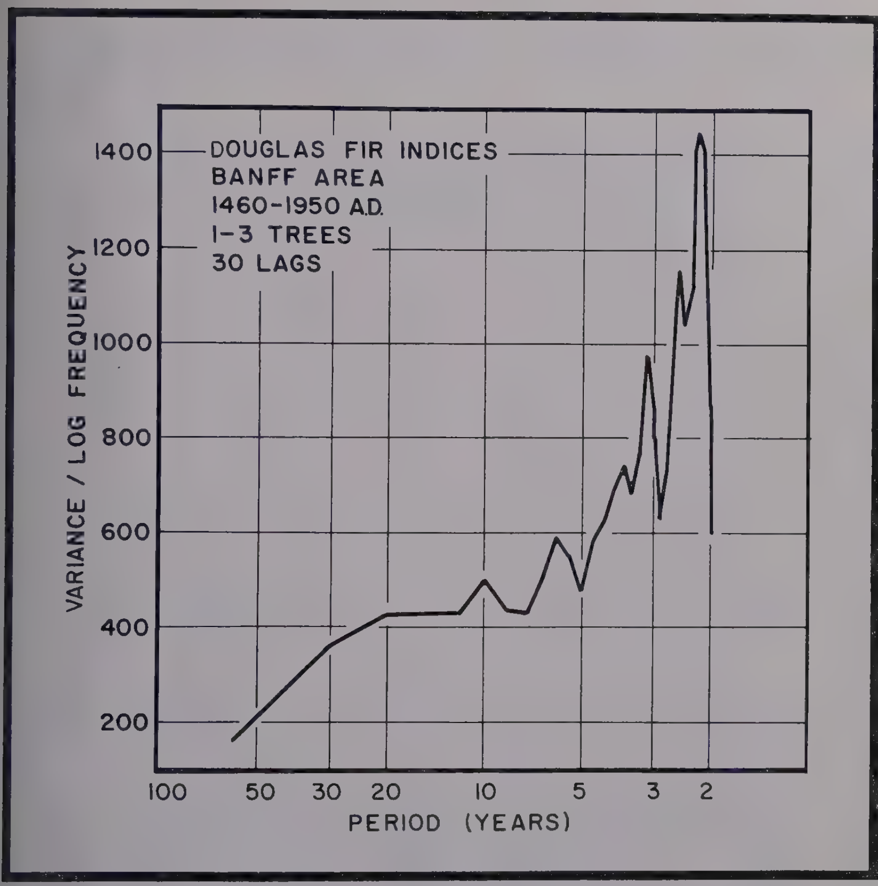


FIGURE 7. The variance spectrum of Douglas fir indices from the Banff area.

The question of sunspot-generated cycles appearing in tree-ring series is one that has often been raised. It has been stated that the 11-year sunspot cycle is to be found in tree-ring series, although not as pronounced as the double-sunspot cycle of 22 to 23 years (Douglass, 1928). There is no question that there is an 11-year sunspot cycle, as FIGURE 13 shows, but subcycles and multiples are not evident. Ignoring this fact, however, we may still examine the tree-ring records to see whether there is a tendency for periodicity of the so-called sunspot frequencies. In this respect, it is well to point out that if there is little variance shown by the tree-ring series at a given frequency, there can be little covariance between tree rings and sunspots at that frequency.

Examination of FIGURES 1 to 12 reveals that none of the tree-ring spectra exhibits significant peaks near 11 years, and as many show minima as show minor maxima. About one third of the spectra show broad peaks from 15 to 30 years that might conceivably be regarded as related in some way to a 22-

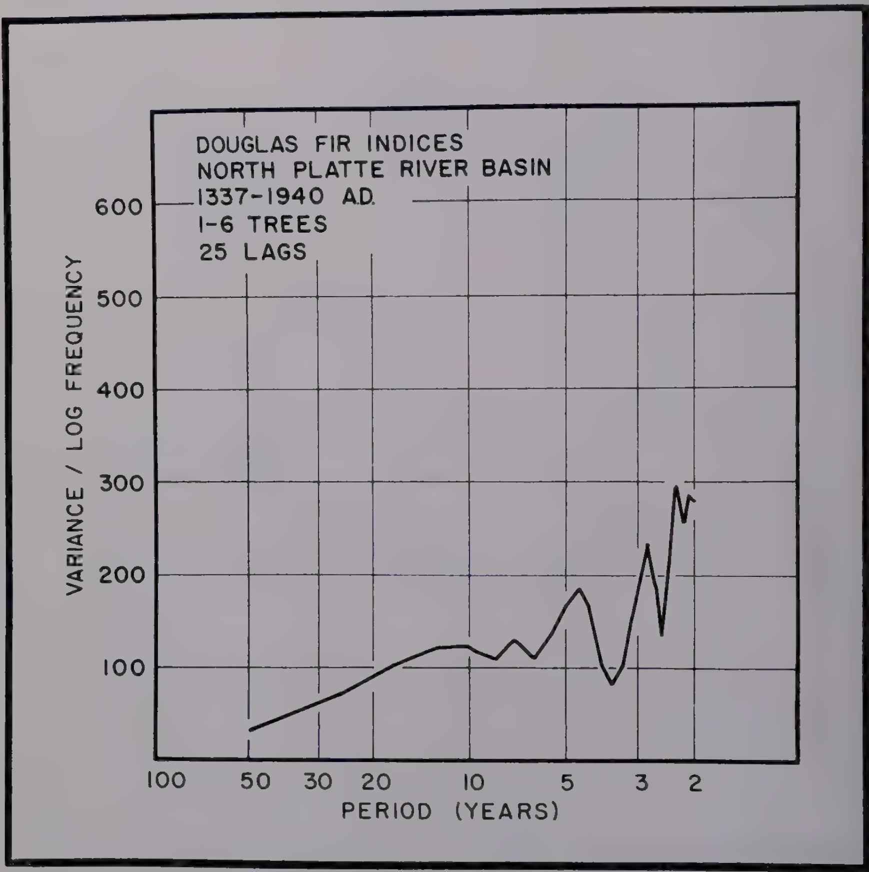


FIGURE 8. The variance spectrum of Douglas fir indices from the North Platte River basin.

year period. Two show completely nonsignificant sharp peaks near 22 years. From this we might conclude that the so-called sunspot periods are not important features of tree-ring spectra. Nevertheless, there still remains a final test of the supposed covariance. FIGURES 14 and 15 present the results of cross-spectrum analysis of tree-ring series versus sunspot series. In each of these figures the variance spectrum of the sunspots is presented as the upper histogram, and the variance spectrum of a tree-ring series for the corresponding

time interval is presented as the lower histogram. Between these in each figure is the plot of coherence as a function of frequency. This displays the distribution of covariance between the two series, as a function of frequency, after accounting for possible phase shifts. The frequencies at which the co-

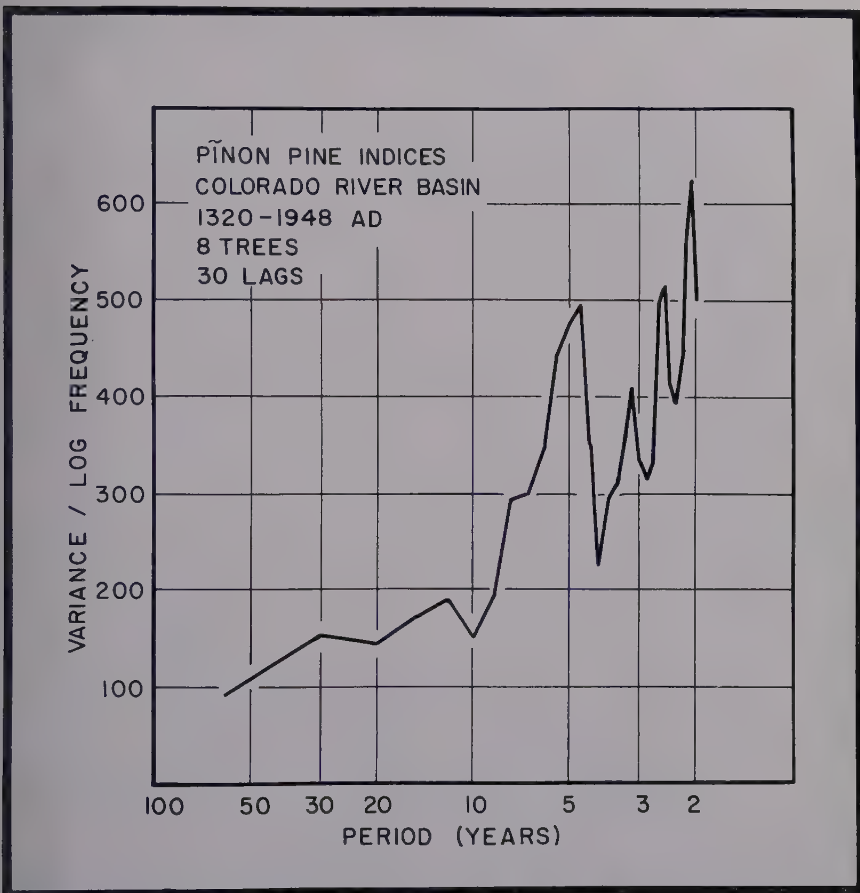


FIGURE 9. The variance spectrum of piñon pine indices from the Colorado River basin.

herence is significant at the 99 per cent level are indicated on the two figures, and the reader will note that these frequencies do not correspond to periods of 5.5, 11, or 22 years. Nor do they agree for the two series, despite the fact that they are from nearby locations.

Summing up the variance-spectrum evidence on "hidden" periodicities we must conclude that they are well hidden, if present at all. Certain short periods seem to be preferred in most tree-ring spectra, and in a spectrum (com-

putational method not given) of July rainfall for the southwestern United States as well (Sellers, 1960). For example, the 2+- and 3+-year periodicities, which appear frequently in the spectra of this study, are also present in the Southwestern July rainfall spectrum. Subjective examination of the spectra indicates the occurrence of tree-ring spectral peaks as shown in TABLE 1.

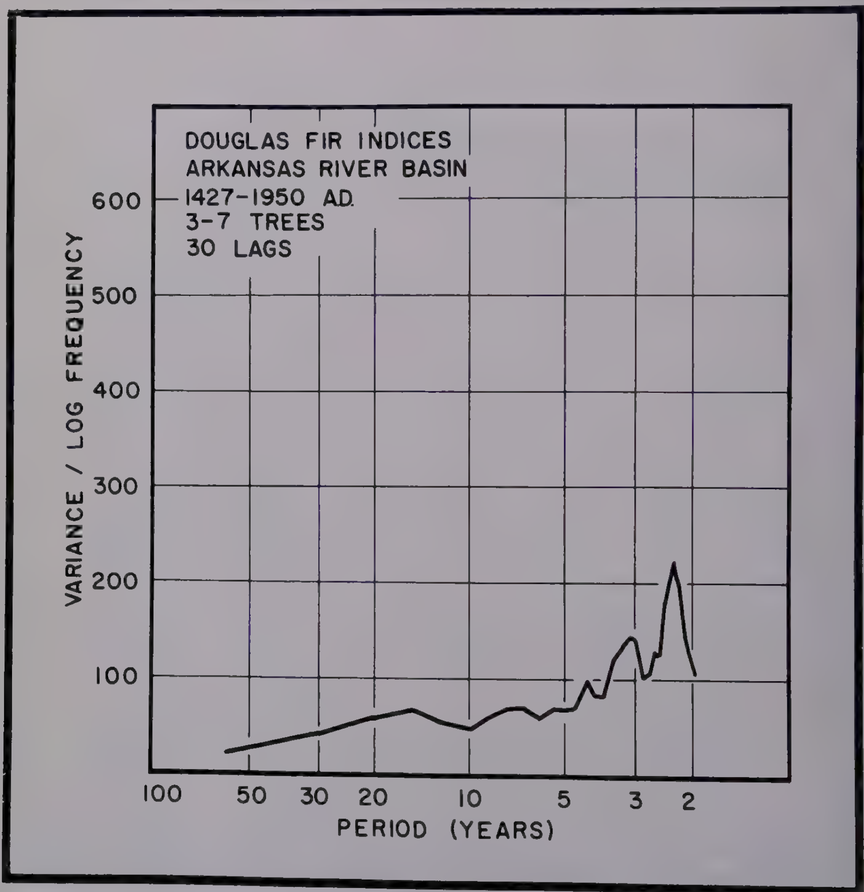


FIGURE 10. The variance spectrum of Douglas fir indices from the Arkansas River basin.

Other than this slight indication of some relatively universal variance excess at 2+ and 3+ years, we must conclude that there are no demonstrably important frequencies in the tree-ring record, and that at higher frequencies the variance is distributed nearly as "white" noise.

Related Problems

Having established that tree-ring series are not characterized by pronounced periodicities, especially not the sunspot periodicities, we still must examine

several questions relating to just what tree-ring series are like. While we have not yet computed the cross-spectrum of the tree-ring and rainfall series, comparison of the spectra presented in this paper with the Southwestern rainfall spectrum of Sellers shows that the humps and peaks are in the right places for strong covariation. In any case, the evidence cited in the literature is

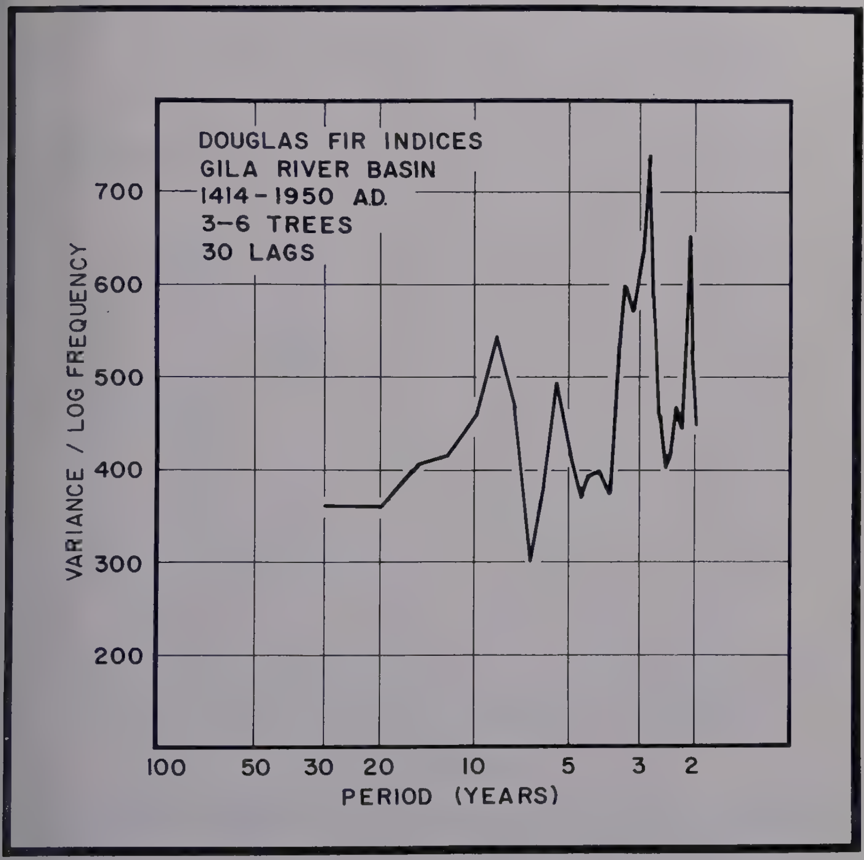


FIGURE 11. The variance spectrum of Douglas fir indices from the Gila River basin.

sufficiently convincing to warrant the conclusion that there is a relation between the two series.

Using the power-law relations already described in this paper, we find that: (1) The sequoias yield generally a minus five-thirds relation between variance per frequency interval and frequency. This implies that the cube of the amplitude of each component Fourier term varies linearly with period. At higher frequencies, the slope is zero, which means that the variance is proportional to frequency, an inherent feature of white noise.

(2) The Snake River limber pine spectrum yields a slope of minus one, as does the Southwestern rainfall spectrum. This means that the variance is constant, independent of frequency. At higher frequencies the limber pine series is nearly white noise.

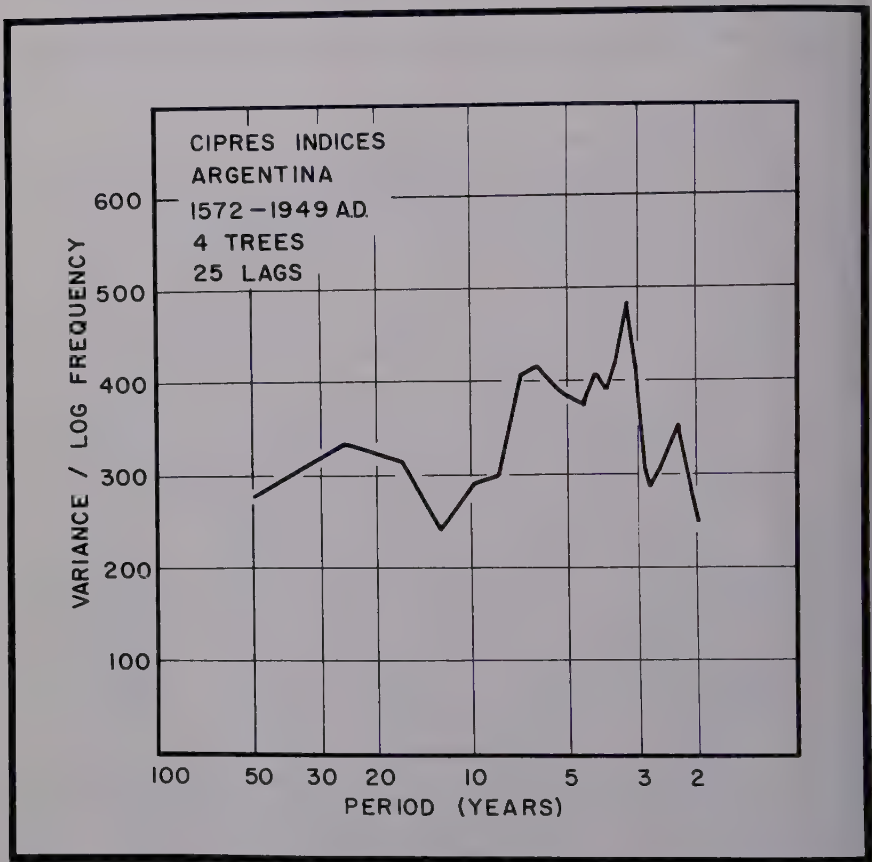


FIGURE 12. The variance spectrum of cipres indices from San Carlos de Bariloch, Argentina.

(3) The spectra of indices have had low frequency variance removed, and thus show white noise at low and high frequencies with variance independent of frequency for periods in the decade range.

It has been suggested that tree-ring series are somehow like varve series or vice versa. The spectrum of a proglacial varve series compiled by Liden is presented in FIGURE 16. It is immediately obvious that the variance is more concentrated at high frequencies than in the tree-ring spectra, and the power-law relation is quite different. Comparing the sequoia spectrum with the

varve spectrum (FIGURE 17), we find that the latter has a slope on the log-log plot of about minus one third rather than minus five thirds or three thirds, as do the tree ring spectra. This means that the variance varies with the two-thirds power of the frequency, an organization unlike that present in the tree-ring series.

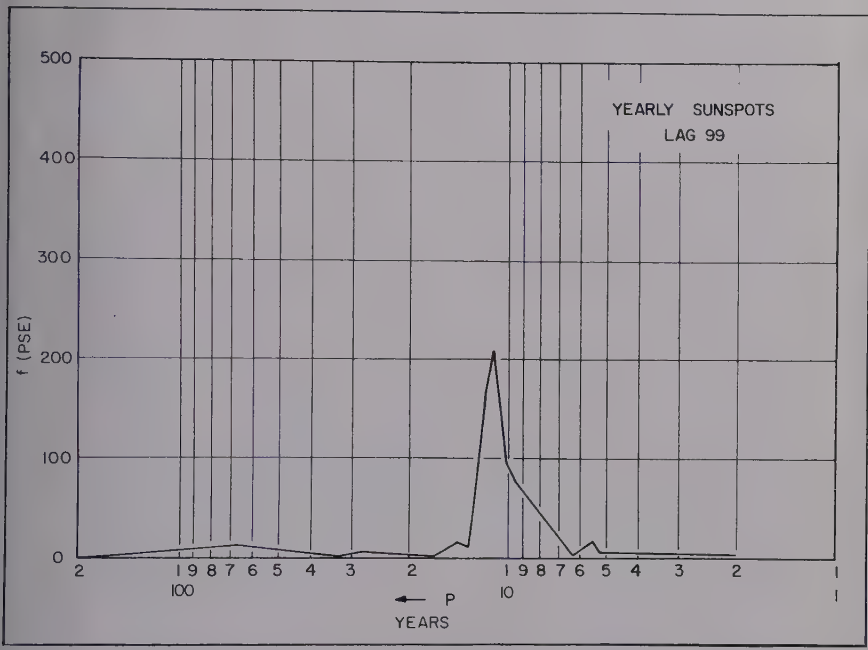


FIGURE 13. The variance spectrum of yearly sunspot numbers from 1760 to the present. The label f(PSE) on the ordinate indicates that the spectral density estimate has been multiplied by the frequency for ease in seeing the spectral details.

Summary

From the results of this study we conclude that:

(1) There is little evidence of *important* periodicities in the tree-ring thickness series that have been studied.

(2) There is a hint of several weak periodicities of widespread distribution, particularly at 2+ and 3+ years, and some evidence of local "ecological" periodicity.

(3) There is no evidence that sunspot periodicities appear in the tree-ring series.

(4) The spectrum of the one varve series examined looks more like the spectrum of a series of random numbers than like the tree-ring spectra.

(5) There is an over-all organization of the variance contained in the tree-ring series, as shown by the power-law relations.

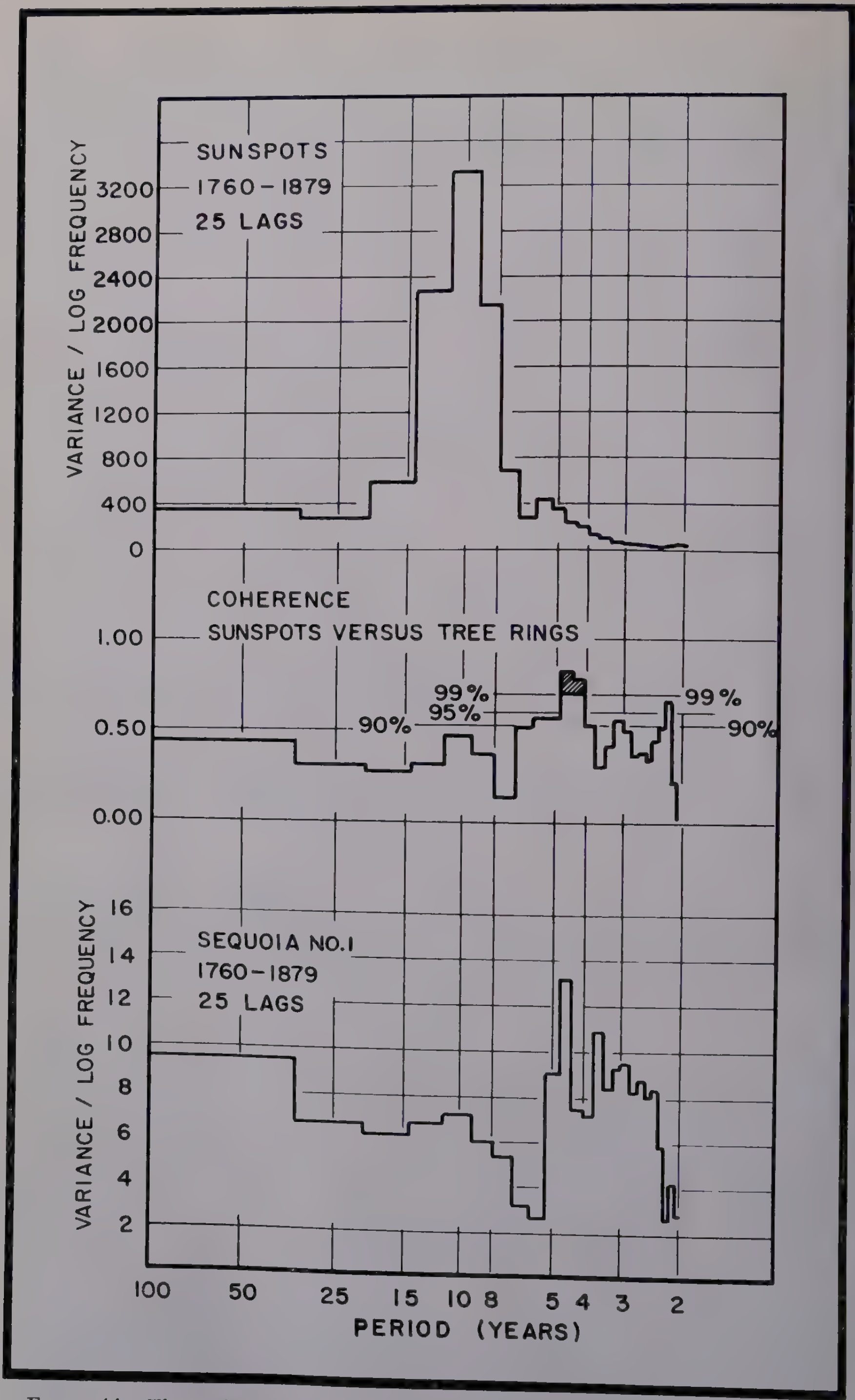


FIGURE 14. The variance spectra of coincident series of yearly sunspot numbers and annual ring thickness from a California sequoia, and the coherence between these series as determined from the cross-spectrum.

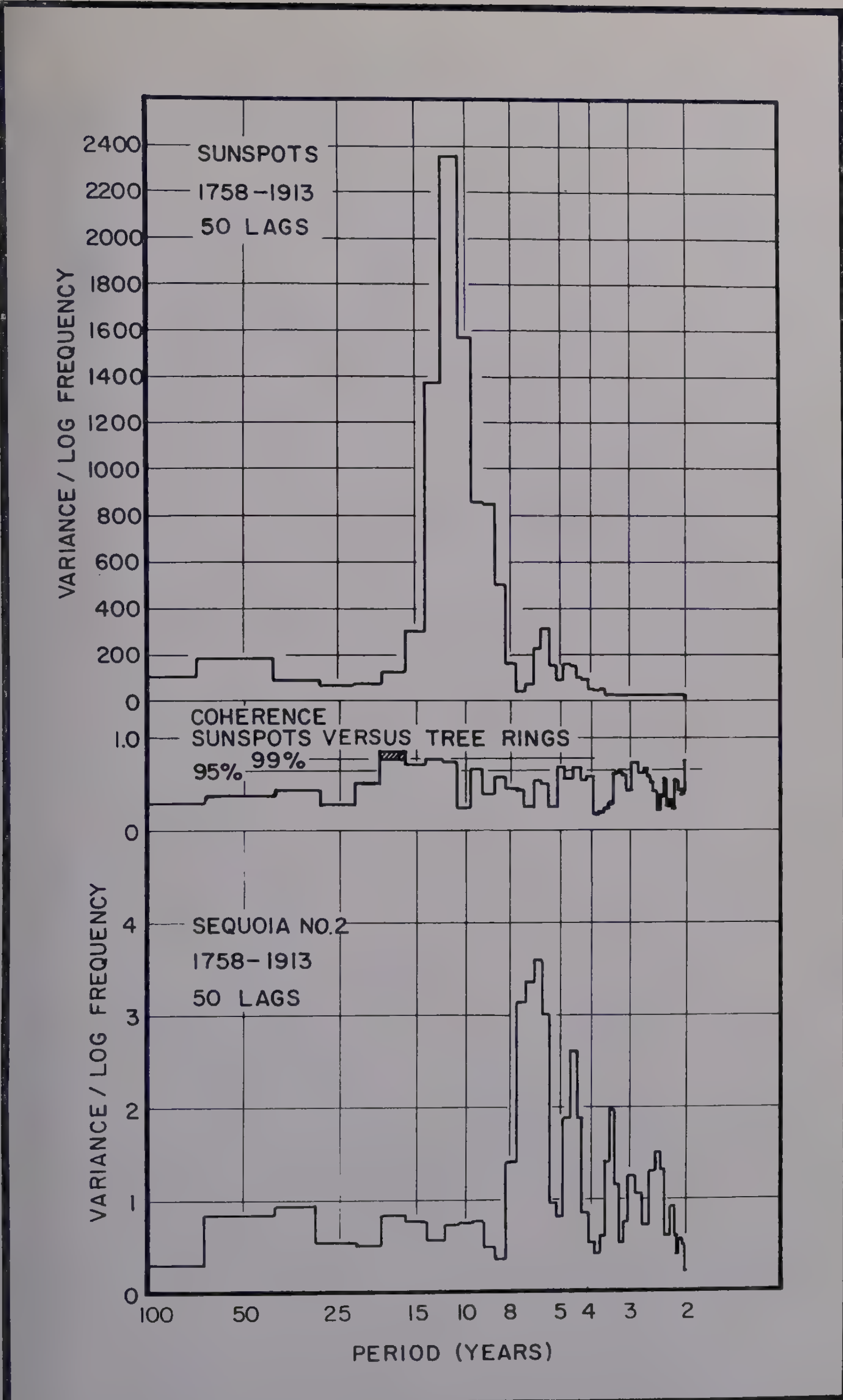


FIGURE 15. The same as FIGURE 14, but using the series of ring thicknesses from a different sequoia.

TABLE 1
TREE-RING SPECTRAL PEAKS

Period (years)	Clear peak	Possible peak	Absent
2.1 to 2.2	12	1	1
2.6 to 2.8	6	3	5
3.1 to 3.3	9	3	2
4 to 5	7	2	5
11	1	5	8
22	0	4	10

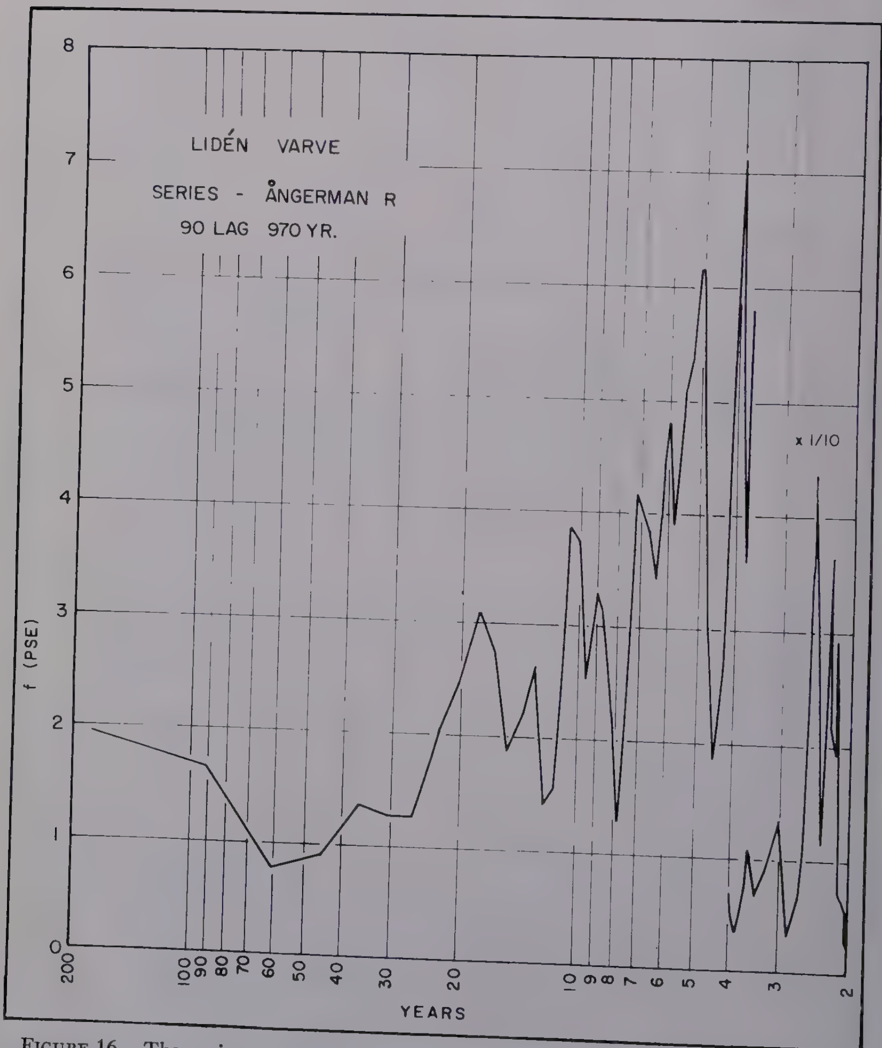


FIGURE 16. The variance spectrum of a 970-year segment of the Lidén varve series from the Ångerman River valley, Sweden. The label f(PSE) on the ordinate indicates that the spectral density estimate has been multiplied by the frequency for ease in seeing the spectral details.

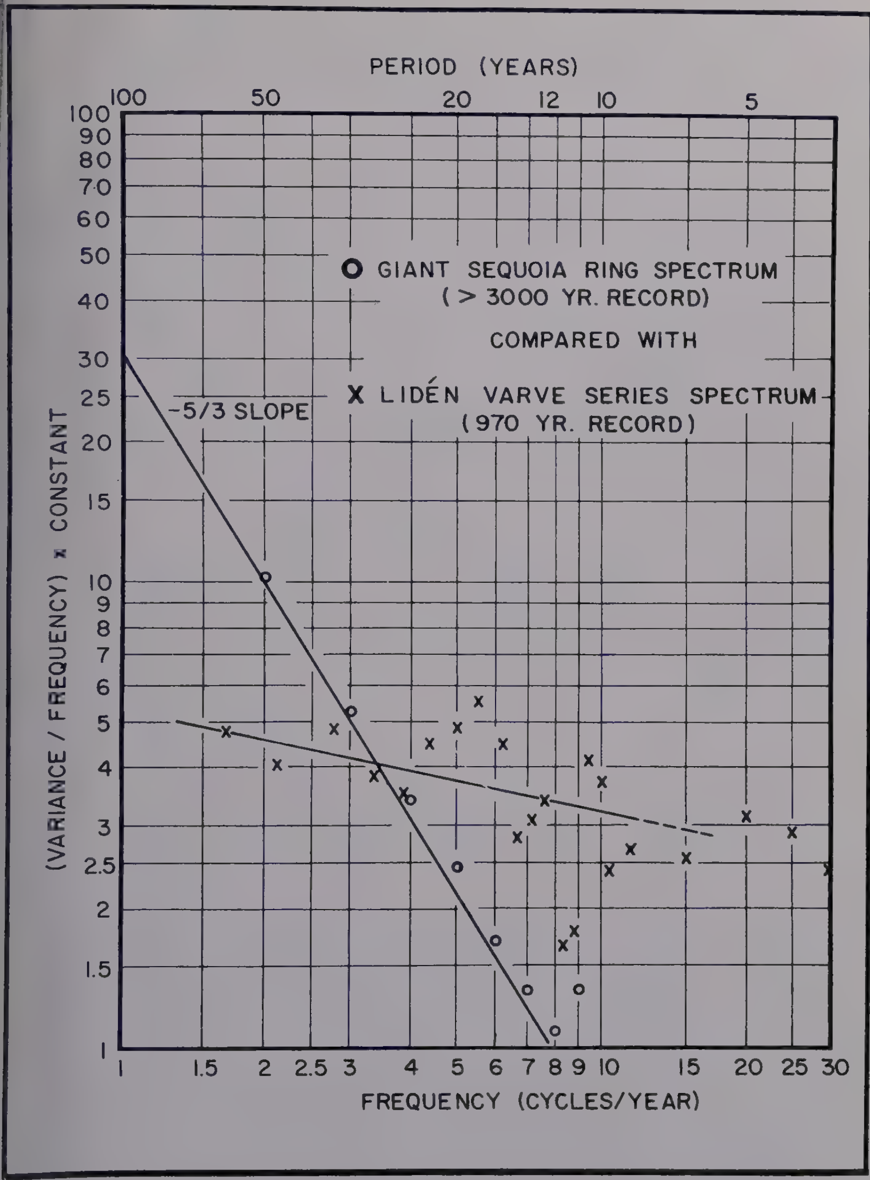


FIGURE 17. Comparative log-log plots of portions of the spectra of a sequoia ring series and the Lidén varve series.

Acknowledgment

Many persons contributed efforts and advice during the preparation of this paper. Of these, we particularly wish to thank Terah L. Smiley of the University of Arizona for data on the series of Miocene tree rings and for a manuscript plot provided in connection with the proglacial varve series presented in FIGURE 16.

References

BLACKMAN, R. B. & J. W. TUKEY. 1958. The Measurement of Power Spectra. Reprinted by Dover Publ. New York, N.Y.

DOUGLASS, A. E. 1928. Climatic cycles and tree-growth. Carnegie Institution Wash. Publ. **289:** (2).

KAHN, A. B. 1957. A generalization of average correlation methods of spectrum analysis. J. Meteorol. **14**(1): 9–17.

LANNING, J. H. & R. H. BATTIN. 1956. Random Processes In Automatic Control. McGraw-Hill. New York, N.Y.

PANOFSKY, H. A. & G. W. BRIER. 1958. Some Application of Statistics to Meteorology. Pennsylvania State University. University Park, Pa.

RUDIN, W. 1953. Principles of Mathematical Analysis. McGraw-Hill. New York, N.Y.

SCHULMAN, E. 1956. Dendroclimatic Changes in Semiarid America. Univ. Ariz. Press. Tucson, Ariz.

SELLERS, W. D. 1960. Precipitation trends in Arizona and New Mexico. Proceedings 28th Annual Western Snow Conference. (April). Santa Fe, N. Mex.

WASSERFALL, K. F. 1930. On periodic variations of terrestial magnetism. Geofys. Publikasjoner. **5:** (3).

WIENER, N. 1930. Generalized harmonic analysis. Acta Mathematica. **55:** 118.

WIENER, N. 1932. The Fourier Integral. : 46. Reprinted by Dover Publications. New York, N. Y.

ZYGMUND, A. 1935. Trigonometrical Series. Reprinted 1955 by Dover Publications. New York, N.Y.

TREE RINGS AND CLIMATIC CHRONOLOGY

Derek J. Schove
St. David's College, Beckenham, Kent, England

Introduction

Tree rings in marginal climates provide indications of rainfall and temperature in individual seasons. In the United States, the practical needs of the water engineer in semiarid regions have justified the emphasis on local rainfall but, from the climatological viewpoint, it is more satisfactory to consider the pressure and pressure pattern over a wide area. Moreover, once the data are sufficiently extensive, past climatic fluctuations can be studied from a global viewpoint. Tree-ring series are accordingly being collected from different parts of the world, and preliminary results (subject to a more exhaustive computer analysis) of the new approach are available for four regions: the subarctic tree line in Scandinavia and Alaska; the semiarid regions of the temperate United States; drought-sensitive timbers of Northwestern Europe; and drought-sensitive timbers in low latitudes.

The Subarctic

A synthesis of North Scandinavian tree-ring series since 1460 A.D. has been published (Schove, 1954*a*). The constituent series showed the usual local variations, but all were affected by one common factor—the summer temperature—so that the combined index reflects the probable warmth of nearly 500 summers. Summer temperature in turn reflects wind and pressure so precisely that the probable pressure pattern of individual summers can be determined directly from the tree-ring index. The typical barometric distribution associated with narrow rings is thus shown by maps to be: low pressure over Finland and high over western Europe; this situation may be described as monsoonal, for it brings northerly winds with much cloud and rain to North Scandinavia. Similar maps for Scottish trees will be published shortly (Schove and Frewer, 1961).

Thirty-year moving means of this index reveal maxima and minima that may be significant. An f(T)-max. *ca. 1573** suggests the period of warm summers (Schove, 1949*a*) that ended suddenly about 1590, and (cf. Flohn, 1957, p. 204) the narrow rings of 1614 to 1650 (f(T)-min. *ca. 1628 or 1630*), 1708 to 1751 (f(T)-min. *ca. 1719*), and 1835 to 1870 (f(T)-min. *ca. 1849 to 1851*) as corresponding to the 3 phases of the Little Ice Age (cf. Table 2;* and C. J. Heusser, this monograph). However, cool summers in this area often synchronize with mild winters in Europe. Indeed tree growth was also low in the period of monsoonal summers from 1902 to 1928 (f(T)-min. *ca. 1913*) when compared with the preceding period of rapid growth (f(T)-max. *ca. 1886*).

Trends in the early part of the original series may, however, be spurious, and here they have been removed by the use of what I term "Relative Anomalies" (Schove, 1953*e*): each decadal value is an anomaly in relation to the enclosing 30-year mean. Thus in Table 1* the value for 1511 to 1520 represents the departure of the tree-ring width from the 1501 to 1530 normal.

* Schove, D. J. Solar Cycles and the Spectrum of Time Since 200 B.C. This monograph.

A similar index is being prepared for Alaska, using tree-ring series between West Alaska and the Mackenzie River, where a cool summer temperature impedes growth. The index given in the above-mentioned Table 1* was a preliminary one of this kind, based on information kindly supplied by Giddings and Oswalt (unpublished; see Schove, 1953*e*, pp. 107–109, 177, and 284) in their articles in *Tree Ring Bulletin*, 1943, 1948, and 1950.

Trends in the original series are inconsistent with those of Karlstrom's series (elsewhere in these pages) and, therefore, they too have been eliminated by the use of relative anomalies in TABLE 1. Nevertheless it is encouraging to find that one f(T)-min., *ca. 1830*, is very near the f(T)-min. evident in Karlstrom's series; the latter series, in so far as it is based on trees at selected sites above timber line, is a model that should be copied elsewhere, for such series should be given extra weight when included in the final index.

Relative anomalies of the Alaskan index suggest that summer warmth in Alaska coincides with summer warmth and dryness in Europe and, probably, also with warmth in the tropics, as there appears to be an inverse correlation with sunspot numbers and with relative decadal anomalies of the pressure parameter.

Semiarid North America

The United States tree-ring data (kindly supplied by Schulman and Smiley) are very accurately standardized, but the search for cycles has been unsuccessful. This is not surprising. Indeed regular cycles—such as one of 11 years—are not the type of oscillation that would be expected on meteorological grounds. In planning cycle analysis it is helpful to know what to look for first and, as far as tree rings are concerned, it is necessary to combine heterogenous series (1) that cover a very wide area, and (2) that are nevertheless affected by a common climatic factor.

There are several types of oscillation that can be studied in United States tree-ring data: (1) annual values associated with the "southern oscillation" of 2 to 3 years; (2) 5-year means associated with an apparent North American oscillation; (3) 5-year means with the pressure-parameter oscillation, including effects ascribed to the 11-year sunspot cycle; (4) 10-year means associated with, for example, fluctuations in the strength of successive sunspot cycles (cf. auroral numbers in Table 1*); (5) 30-year means reflecting major pressure oscillations (for example, 60 to 70 years); (6) major temperature oscillations (for example, 80 or 90 years); and (7) major solar variations (for example, the 200-year cycle).

The southern oscillation. Barometric pressures for the Indian Ocean are available since 1796 (Schove and Berlage, in preparation), and the data collected by Schulman (1956) show correlations that may at first seem surprising. Rainfall in many Pacific regions thus fluctuates positively with these annual pressure values. Tucson, Ariz., and Santa Fe, N.Mex., tend to follow this "Pacific" pattern. The United States is in a transition zone and the opposite or "Indian" pattern is followed by the rainfall of nearby Colorado and upper Colorado and by the upper Missouri River runoff in Montana. A

* Schove, D. J. Solar Cycles and the Spectrum of Time Since 200 B.C. This monograph.

similar contrast in tree-ring reactions can be noticed between Pacific sites to the south (for example, east-central California, west-central Mexico, Mexico, and Big Bend, Tex.) and Indian sites east of the Rockies (Banff, Alta., Canada, and the Missouri River basin). In some cases there is a phase difference of as much as 12 months. All these relationships are nevertheless weak and unreliable, and North American tree rings have not so far been included in our attempts to study the southern oscillation in past centuries.

A negative correlation of certain United States tree-ring data with the Scandinavian summer temperature index (Schove, 1954*a*) has been announced by Cross (1961), who is measuring the effects of snow pack in different years on the shape of tree rings at Mount Rainier, Wash.

Correlations of annual tree-ring values with N-S pressure differences or the north Pacific oscillation in different seasons should prove most illuminating but, as far as I am aware, this has not yet been attempted. Likewise, qualitative features (cf. Schove and Lowther, 1957, p. 89) should prove significant.

North American oscillation. There is an oscillation in 5-year moving means of the climatic elements of the United States that might seem to be some North American oscillation. During the period 1901 to 1930 its wave length coincided with that of the sunspot cycle, which therefore appeared to be a convincing cause of the fluctuations shown by charts of moving 5-year means (for example, as in Tannehill, 1947). The most obvious effect of this oscillation is the negative correlation between air (and sea) temperatures on the Pacific coast, on the one hand, and the air (and sea) temperatures for the remainder of the United States on the other. Pressure is higher on the Pacific coast when it is colder in the southwest, but this pressure anomaly is opposed to that in the tropics of the Old World.

After 1920 the sea-temperature information published by Rodewald (1956) leads to the following differential values for the Atlantic minus the Pacific coasts:

1920 to 1924 ca.	−1
1925 to 1929 ca.	−11
1930 to 1934 ca.	−1
1935 to 1939 ca.	+1
1940 to 1944 ca.	−8
1945 to 1949 ca.	+8
1950 to 1954 ca.	+13

(Temperature is in degrees centigrade multiplied by 10, expressed as departures from 1925 to 1954 normals.) Presumably these values will be found to match well with temperatures in the United States (cf. Mitchell, this monograph) as a whole, and with pressure off the Pacific coast. The main peaks are the 5-year periods centered at 1881, 1894, 1910, 1922, 1937, and 1952(?) with the intervening minima at 1886, 1905, 1916, 1927, and 1941.

Table 1, based on curves in Tannehill (1947), illustrates the patterns at a time when the cycle was about 10 years. This oscillation arises from differences in pressure pattern often associated with the sunspot cycle initially through what is termed a world pressure parameter. There is therefore no

separate North American oscillation, but 5-year means of responsive American tree rings may help us to determine pre-1796 values of this important index discussed below.

There is another cycle that affects Greenland, Canada, and the northeastern part of the United States, and this seems to be an oscillation in the upper air trough in the westerlies over the Davis Strait. Five-year moving means indicate pressure maxima (1878, 1887, 1893 . . . 1908, 1917, 1925, 1932/1933, 1941, 1948, and 1956 in W. Greenland) that reflect the parameter changes discussed below and that also indicate a cycle that changed from 7 to 8 years about 1900 (when the general pressure fell).

The world pressure parameter. World pressure maps of changes by 5-year periods have been constructed from barometric information received as a result of international appeals (cf. WMO Bulletin, January, 1959, p. 40). Changes can be grouped into two opposing types, according to the sign of the pressure change in west Greenland. A change of opposite sign but smaller

TABLE 1

CENTRAL DATES OF MOVING FIVE-YEAR MEANS OF CLIMATIC ELEMENTS IN THE UNITED STATES

Sunspot maxima (actual year)		1883	1894	1907	1917
Pacific coast pressure	Max.	1881/1882	1893	1910	1920/1923
Pacific coast temperature	Min.	1881	1894	1910	1920/1921
United States (especially the East) temperature	Max.	1880	1898	ca. 1910	1920/1923
Pacific coast pressure	Min.	1886	1905	1915	1926
Pacific coast temperature	Max.	1888	1904/1905	1915/1916	1926
United States (especially the East) temperature	Min.	1886	1904	1917/1918	1928
Sunspot minima (actual year)		1889	1901	1913	1923

magnitude occurs simultaneously in the Indian Ocean where mean pressures can be accurately determined for over a century. Indeed, at present, it is convenient to use the reversed mean of several areas in the neighborhood of the Indian Ocean as the basis of what is termed the "first approximation to the pressure parameter". Successive approximations of increasing complexity will be made, but this is already a close approximation, and values are given in TABLE 2.

The similarity with the southern oscillation is thus obvious. However in the ordinary southern oscillation, as Berlage (this volume) has made clear, there is little correlation with the pressure in Greenland that, if anything, in individual years tends to agree with the Indian Ocean rather than the Pacific. In the case of these 5-year changes it now appears as if the southern oscillation is geared together with the North Atlantic oscillation. Moreover this oscillation is rather more extensive: the term Indian Ocean has now expanded to include the Cape of Good Hope and southeast Australia!

The "signature" of the pressure parameter during the 9 lustra of the period 1871 to 1915 is given by the mnemonic code sequence P, O, N, O, P, O, N, P,

N where P and N refer to relative peaks—positive and negative respectively—and O to those 5-year periods such as 1876/1880 that were transitional. This suggests an oscillation of decreasing wave length, and does not always agree with the sunspot cycle. This pattern is followed for: (1) reversed pressure

TABLE 2
THE PRESSURE PARAMETER AND FIVE-YEAR MEANS OF SIGNIFICANT CLIMATIC INDICES

Lustrum*	Reversed Indian Ocean pressure mb. × 100†	Signature‡	Sign change from previous lustrum§	Tropical temperature CX 100 ‖	Sunspot turning points¶
1823	−?			(+)	N
1828	+?		+	(−)	P
1833	−?		−	(+10)	N
1838	+?		+	(−20)	P
1843	(−)		−	(+10)	N
1848	(+)		+	(−10)	P
1853	(+4)		−	(+10)	N
1858	+9		0	(−0)	P
1863	+19		+	(−20)	O
1868	−9		−	(+10)	N
1873	+9	P	+	−2	P
1878	−15	O	−	+11	N
1883	−21	N	−	−14	P
1888	−12	O	+	−9	N
1893	+9	P	+	−15	P
1898	+1	O	−	+15	O
1903	−11	N	−	−1	N
1908	+7	P	+	−11	P
1913	−13	N	−	+9	N
1918	+11	P	+	−12	P
1923	+13	or P	0	−1	N
1928	−4	N	−	+15	P
1933	+7	P	+	+9	N
1938	+5	or P	(0)	+15	P
1943			(−)	+21	N
1948			(+)	+21	P
1953			−	+	N
1958				+	P

* Each lustrum is denoted by its middle year, for example, 1843 = 1841 to 1845.

† First approximation to pressure parameter (relative to 1881 to 1940 normals); the relative maximum in 1921 to 1925 is not representative of better approximations.

‡ Mnemonic signature of relative maxima (*P*) and minima (*N*).

§ For calculation of agreement coefficients. Values after 1935 are subject to correction when, for example, Greenland pressures become available.

‖ Date from Callendar (1961) relative to 1901 to 1930 normals. Estimates before 1870 are partly based on additional information kindly supplied by Callendar.

¶ Relative maxima and minima by lustra.

over the Indian Ocean and neighboring areas; (2) rainfall for regions of Indian type, and therefore the Nile flood area; (3) temperature in the tropics generally, with the notable exception of 1881/1885 (Krakatoa?); (4) pressure in southwest Greenland (except for 1871/1875), in temperate South America, and probably the Pacific; (5) ice in Davis Strait less that in Iceland and Newfoundland, except for the periods 1881/1900; (6) rainfall reversed in the south of the United States between the Rockies and the Mississippi River (also many

South American regions for most of the periods); and (7) sunspots, but not between 1876 and 1890 (cf. also divergence after 1920).

With such an extensive list of correlations, it will be a simple matter of arithmetic to combine tree-ring series from different parts of the world to evaluate the values of this parameter back at least to the years 1 to 5 A.D.! The principle of selection could be the agreement coefficient (cf. Schove and Lowther, 1957) between successive 5-year changes.

Many vague effects hitherto ascribed to the 11-year sunspot cycle show a much better relationship with this world pressure parameter, which presumably reflects more exactly than sunspot numbers the effective solar radiation. It is unlikely that tree-ring widths by themselves could be used to correct any errors in my dates of historical maxima (cf. Schove, 1955*b*), but it will be interesting to watch for divergencies such as have occurred since 1920 between the sunspot and parameter maxima.

Five-year means of tree-ring changes in North America again fall into two groups, but some series where the annual data were of Pacific-type are now of Indian-type, that is, tree-growth over 5 years is greater when Indian Ocean pressure is lower. Nevertheless the marginal location of the region has again made it advisable to omit both series in investigating the pressure parameter itself. This marginal location also explains why the sunspot cycle rarely shows up in United States trees.

The sunspot cycle does appear in some United States trees if my auroral minima (Schove, 1955*b*) are used instead of the rigid 11.4-year sunspot cycle conceived by Douglass. The central Pueblo series described by Douglass (1936, Vol. 111: 106, 107) reflects the cycle in much the same way as a rainfall series of Pacific-type, but with about 2 years delay between sunspot minima and tree-ring maxima. Presumably the apparent sunspot effect is again indirect, and the differences between United States trees of the two types might indicate not so much the sunspot cycle but the pressure parameter fluctuations as reflected in a North American oscillation.

Ten-year means and auroral numbers. The sunspot cycle is almost eliminated if overlapping 10- or 11-year means are used. It is surprising therefore to find that sunspot numbers are now more and not less important. At the time when the auroral numbers of Table 1* were prepared, it did not seem likely to me that meteorological correlations would be significant.

The parameter, not the auroral numbers, should again be regarded as the immediate factor necessary to explain climatic anomalies and, although Faegri's rule is not true in the contrast between 5- and 10-year changes, the maps of the changes fall roughly into the same two phases of the fundamental "standing wave" of southern-oscillation type. A first approximation to this parameter is provided by the tree-ring indices of Table 1,* column E (cf. also column D).

The effects of these changes of pressure pattern have sometimes been noted and ascribed to other causes. Thus from 1880 to 1945 there was an accidental 22-year cycle that appears as an attractive scapegoat. However, a number of similar curves from different parts of the world assembled by Yamamoto (1957), although ascribed by other climatologists to coincidence, do fit well the patterns anticipated.

* Schove, D. J. Solar Cycles and the Spectrum of Time Since 200 B.C. This monograph.

The signature of the 10-year pattern is, in the instrumental period, best shown in decades of the form 06-15, for then the mnemonic for successive decades from ca. 1870 to ca. 1930 reads PNPNOPN, although the period 1866 to 1875 is not a period of minimum pressure in the Indian Ocean (cf. TABLE 2) itself. That sequence is nevertheless characteristic of both climatic anomalies and sunspot changes, and it provides an objective criterion when tree-ring data are used to determine the decadal pressure parameter in past centuries.

Once again a high parameter normally means wetness at Indian-type stations and dryness at Pacific-type stations. This time, however, most good tree-ring series, from North Platte, Nebr., to Arkansas and across the Rockies to Colorado and even southern California (but not, for example, Tioga Pass) are—in the period of the signature—of Indian type. A column (D) has been included in TABLE 1* to represent the mean growth over this region but, as all the trees concerned are within the United States, anomalies peculiar to the United States also play their part. This explains why the disagreement percentage by lustra with an Argentine series of Pacific-type trees (40 per cent) was less instead of more than 50 per cent.

It will be possible to obtain a good index of the parameter from tree-ring series in the neighborhood of the Indian Ocean, but the values of Table 1,* column E, are included to illustrate the method rather than provide the results, as they are based on the only three sets of standardized series available at the time of writing.

The North American series provide meanwhile a useful approximation to the parameter, for the use of running 10-year means reveals the same sequence of t-max. and t-min. as found thermometrically and as mentioned above. Moreover these warm-dry and cold-moist waves are now seen to be the phases of pressure parameter waves associated with decadal sunspot minima and maxima respectively. The dryness of the dry phase extends from the old-world tropics to North America as far north as Montreal, P.Q., Canada.

Low sunspot numbers about 1810 thus fit with a warm wave evident in the parameter and in the Alaskan trees and a "long dry" period (noted below) in the United States.

Sunspots are still relatively low about 1830 when the next low-parameter pattern appears. Again trees in Alaska and in the Rockies reflect a warm-dry wave dated instrumentally to ca. 1829/1830. The rapid rise to an s-max. ca. 1835/1840 is, however, associated with the usual high parameter pattern and cold wave. In northern Italy the t-min. is dated again rather early (ca. 1833), but in the United States the instrumental date of ca. 1839 is perhaps a little late in relation to the characteristic Japanese famines of 1833 to 1839.

The next warm and cold waves occur in rapid succession in the United States about 1865 and 1870 respectively. These dates do not fit the parameter approximation given in TABLE 2, but the sudden revival of sunspot activity from s-min. to s-max. in the 1860s resembled that of the 1830s and seems to have given similar effects (for example, in Japan, as noted by Arakawa, 1955 and 1960, p. 112).

The warm wave of 1876 to 1881 followed the sudden collapse of sunspot activity and is this time reflected in a very considerable decrease in the pressure

* Schove, D. J. Solar Cycles and the Spectrum of Time Since 200 B.C. This monograph.

parameter associated with droughts in several continents. The cold wave followed about 1888/1890, ahead of the parameter and the s-max.; it was possibly partly induced by the eruption of Krakatoa, in the East Indies, and it extended to Central Europe.

About 1900, there was the most striking s-min. of recent times, again with widespread subtropical droughts. There was, at the same time, a well-marked minimum in the various effects of the parameter change. The pressure pattern change was described by me (Schove, in Kraus, 1956, p. 358) before I appreciated the part played by sunspots. The warm, dry phase occurred about 1898, both in the tropics and in the United States, but it was not until ca. 1920 that a further high value occurred of either the s-max. or of the pressure parameter.

A p-min. about 1930 fits a minimum in the parameter and a further warm wave, although this is less clearly defined. The t-max. ca. 1934 is nevertheless common to the eastern United States and to Central Europe.

Although subsequent waves also are indefinite (cf. transatlantic t-min. ca. 1938, t-max. ca. 1948) the inevitable fall from the present high sunspot numbers will presumably be associated with a sudden fall in the parameter, and a further warm, dry wave extending from the old-world tropics to the United States may yet be associated with drought famines, as in 1877/1878.

The relations between the various indices of Table 1* will be discussed in a forthcoming study of the history of European climate.

Thirty-year trends, the major pressure oscillation ca. 1890 to 1920. Homogeneous pressure series for different parts of the world have been prepared since my FIGURE 3* was first published and it is becoming possible to determine the dates of the P-max. and P-min. in low latitudes. It is now clear that south steering is not important south of 40° north but that, instead, the dates assembled indicate a major pressure oscillation, a seesaw between the Pacific and the Indian Oceans that resembles the southern oscillation. It is interesting to find that "superoscillation" has a different wave length from the major temperature oscillation dominant in the west Greenland center of action.

The pattern is thus:

P-min. Pacific Ocean ca. 1890 to 1891	P-max. Pacific Ocean ca. 1920
P-max. Indian Oceanca. 1890 to 1891	P-min. Indian Ocean ca. 1923

The Pacific P-min. *ca. 1890 to 1891* included the United States Southwest and even Argentina. Simultaneously the Indian P-max. extended from southeastern Australia to Syria, and reinforced the south-steering high that was migrating from central Norway *ca. 1889* to Austria *ca. 1894.* Standard deviations of most climatological elements were reduced even in Europe at this time. In the west, the Indian area probably extended across Africa to Brazil and the southeastern states of the United States, but uncertain barometric records make it difficult to be specific about this. The nodal zone from the Rockies to perhaps southwestern Brazil would seem to have been a narrow belt with winds that were appreciably anomalous in relation to 20th-century normals, and rainfall distributions must also have been peculiar.

A P-max. *ca. 1919/1922* affected much of the Pacific extending from Welling-

* Schove, D. J. Solar Cycles and the Spectrum of Time Since 200 B.C. This monograph.

ton, N.Z. and, once again, to the United States Southwest and across South America to the Argentine.

Almost simultaneously, the P-min. of *ca. 1923/1925* affected the same area as the p-max. of *ca. 1890/1891*. Thus it extended from the southeastern states of the United States (*ca. 1923/1925*) to parts of northwestern Europe on the one hand, and from the subtropical belt to Portugal and Syria on the other. India, Java, and even Hong Kong are also included. In the Southern Hemisphere, it extended from southeastern Australia west-north-west and, probably, across Africa to Northeastern Brazil.

The P-max. and P-min. often correspond to an R-min. and R-max., but there are exceptions, often within short distances of one another. Pacific-type reactions occur in the northwestern United States (cf. *Monthly Weather Review* **63**, pp. 19–23) but, in the Rockies, Indian reactions are more common. Tree-ring maxima and minima faithfully reflect these differing responses and, in the fully standardized series of Schulman (1956) the R-max. and the f(R)-max. often agree to the exact year.

An inverse correlation exists between 30-year curves of tree growth in Pacific areas such as Banff and Jasper, Alta., Canada, on the one hand and the southern and eastern sites (for example, southern California, Rio Grande, and Colorado) on the other. Let us suppose that differences between the two areas reflect major pressure oscillations of the same geographical pattern as that of the half-wave *ca. 1890* to *ca. 1920*. We can then determine the dates of barometric maxima and minima in regions of Pacific-type as follows:

ca. 1530	P-min.	Important	(SW in Europe)
ca. 1560	P-max.	Important	(NE in Europe)
ca. 1590	P-min.	Important	
ca. 1635/1645	P-max.	Important	
ca. 1670	Minor p-min.		
ca. 1685	P-max.		
ca. 1710	Important p-min.		
ca. 1735	P-max.		
ca. 1760	Irregular p-min.	Minor	(SW in northwestern Europe)
ca. 1790	P-max.		
ca. 1815	P-min.		
ca. 1840	P-max.	(NE in northwestern Europe)	
ca. 1885	P-min.	(Model)	(NE in northwestern Europe)
ca. 1920	P-max.	(Model)	(SW in northwestern Europe)

These results are evidently significant in United States climatic history but, until an index can be found that correlates inversely with an Indian Ocean tree-ring index, we must suppose that a complex function of more than one parameter is involved. In particular, we must note: (1) there is no inverse correlation with Table 1* column E (Java minus Pacific); (2) there is no obvious correlation with sunspot numbers; there was an S-min. ca. 1890, but it lasted until *ca. 1910* and there was no S-max. *ca. 1920* to *1925*; (3) wind variations in Europe (indicated from Figures 2 and 6* by the bracketed figures) are independent; (4) there is no regular cycle, such as a Brückner cycle of 35 years; and (5) another parameter implicated in the above dates is the major temperature oscillation that will be discussed below.

Thirty-year means, the major temperature oscillation. The temperature

* Schove, D. J. Solar Cycles and the Spectrum of Time Since 200 B.C. This monograph.

changes, for example in Greenland, reflect another climatic oscillation with a wave length—in the past 2 centuries—of the order of 80 years, thus comparable (although out of phase) with the long sunspot cycle.

The climatic amelioration that set in suddenly in the subarctic about 1895 is associated with climatic changes elsewhere, such as are indicated by Kraus (in these pages), who notes the rainfall decrease on the eastern side of continents. A similar drying (Schulman, 1956, Figure 14) in tree-ring areas of the Pacific slope in western states can also be noted in TABLE 3.

The signature of a single wave is insufficient for us to use, for example, for lakes in Oregon or tree rings in British Columbia, as inverted thermometers, despite the excellent correlation (cf. Shutler *et al.*, this monograph) of pluvials and glacials in the past. Nevertheless if, for the sake of argument, we use the same series of trees to date the phases of the Little Ice Age, we would have: Cold I *ca. 1605*, Warm *ca. 1630*, Cold II *ca. 1675*, *ca. 1715* and *ca. 1750*, Warm *ca. 1780*, and Cold III *ca. 1820*.

TABLE 3
THE MAJOR TEMPERATURE OSCILLATION

	Ice		Trees	Rain	
	Davis Strait	Iceland	Pacific slope		Eastern United States
	Speerschneider	Koch, Schell, Lamb	Schulman		Lysgaard
f(T)-max.	*ca. 1854 or 1841*	*ca. 1850*	*ca. 1845*	R-min.	*ca. 1850*
f(T)-min.	*ca. 1872*	*ca. 1878*	*ca. 1875*	R-max.	*ca. 1875*
f(T)-max.	*ca. 1930*	*ca. 1933*	*ca. 1930*	R-min.	*ca. 1920*

Comparisons between the two half centuries 1851 to 1900 and 1901 to 1950, such as are made by Schell (in these pages) for the year as a whole, are of special interest in this connection. Preliminary results (received in response to my appeal) for the 12 separate months suggest that, although the westerlies have become stronger in all the winter months (November to April) in Europe, the trend has been toward an increased northerliness in summer months (June to October), as if monsoon effects were more significant and solar radiation more effective in the period 1901 to 1950.

Long cycles since ca. 25 B.C.: the 200-year solar cycle. Tree-ring data in the Southwest enable many wet and dry phases to be identified before 1100. Well-marked f(R)-max. can be dated *ca. 25* B.C., *ca. 194* A.D., *ca. 545 to 580*, *ca. 900*, and *ca. 1020*, for example, and well-marked f(R)-min. *A.D. 27*, *ca. 250*, *ca. 770* and *ca. 970 to 1000*. Lack of evidence prevents us from determining the dates of S-max. and S-min. in some cases, but there is again no obvious correspondence and indeed no obvious periodicity. The 200-year cycle in sunspot activity (Schove, 1955*b*, pp. 143, 144; Dewey, 1960) is possibly matched in a drought cycle noted by Schulman between 1200 and 1500, but the latter oscillation may well be accidental; we must await standardized series from the

4000-year-old trees now being investigated before such long cycles can be determined from tree-ring analysis.

Drought-Sensitive Timbers of Northwestern Europe Since 100 B.C.

Oak trees in northwestern Europe are not as long-lived as United States trees but, at most sites even in a normal English summer, they do not get sufficient water for maximum growth, so that the rings are drought-sensitive. Archaeological timber from the Roman and Anglo-Saxon periods can be dated approximately (Schove, 1959, pp. 288–290). Later medieval timber is securely dated, and the narrow rings are often explained by droughts recorded in the chronicles. In the New World, the only comparable evidence is in the Aztec records of Mexico. However, old-world chronicles are not always reliable, and the tree-ring evidence in one instance led (Schove and Lowther, 1957) to the discovery that a date (1231) in a medieval printed chronicle was incorrect; in another it threw doubt on the 7th-century drought of St. Wilfrid, a drought used by C. E. P. Brooks as evidence for his belief that the 7th century was the driest phase of the Dark Ages. Meanwhile, meteorological extracts from unpublished manorial rolls have now been published and found (Titow, 1960, p. 365) to confirm the tree-ring dating of droughts in the 13th century.

The English oak trees show cycles that are probably physiological and are certainly not of solar origin. They also show intriguing climatic fluctuations. In the Roman period, for instance, drought-sensitive trees in southeastern England show minima: termed f(R)-min. (cf. above) *ca. 100 B.C.*, *ca. 100 A.D.*, *ca. 160*, *ca. 205*, *ca. 255*, and f(R)-max. *ca. 40 B.C.*, *A.D. 135*, *185 and 225* (Lowther's measurements and approximate dating on archaeological grounds). Tree-growth over 30-year periods probably depends on, for example, temperature as well as annual rainfall but, with the aid of the "British Tree-Ring Project," it is hoped to prepare a standardized series and to discover what the other factors are. In the meantime we must be cautious about similarities between minima such as 1246 to 1290 that are common (cf. FIGURE 1) to English oak and to trees in the United States semiarid zone.

Low Latitudes

Drought-sensitive trees in low latitudes are especially useful, but few standardized series are available. The pioneer work of Berlage (this volume) in Java is only now being followed up in southwestern Asia and Africa. Individual series show little indication of the sunspot cycle, but they do show very clear influence of tropical pressure patterns. The Java series, compared with pressure maps (unpublished) can be shown to reflect the southern oscillation in general and pressure in Australia in particular. By summation of different series—allowing for phase lags where necessary—it is hoped to obtain an index of Indian Ocean pressure (and thus the southern oscillation) extending back many centuries. Comparison with the dates of sunspot minimum (Schove, 1955*b*) will then make it possible to investigate further the rules indicated by Berlage (in these pages) and Visser (1959). The decadal values in Table 1* in the column termed "Java minus Pacific," although unrepresentative at present, show how an index could be constructed to reflect the pressure parameter and the major pressure oscillation.

* Schove, D. J. Solar Cycles and the Spectrum of Time Since 200 B.C. This monograph.

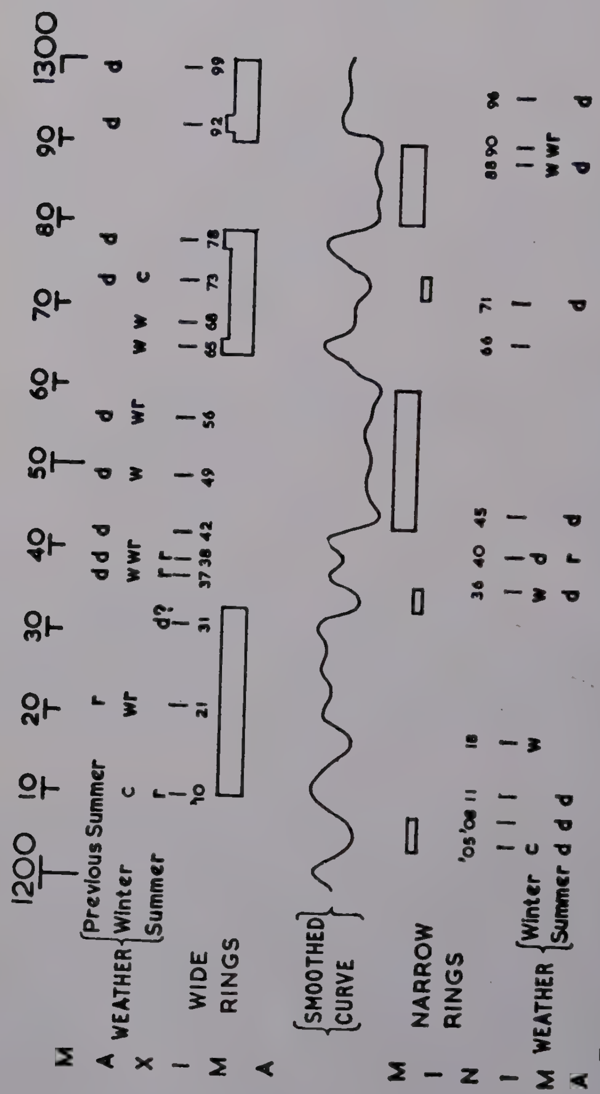


FIGURE 1. Skeleton plot of the 13th century. Oak found at Westminister or in Hampshire, England. Reproduced by permission of *Medieval Archaeology* (Schove and Lowther, 1957, p. 83).

In the concluding section, a survey will be made of the last 185 years' weather in the United States in the light of the principles described in my two papers in this monograph.

Climatic Patterns in the United States Since 1776

The year 1776 happens to mark an early stage of a long dry period in the semiarid parts of North America. It is also a convenient date to begin our review because, after the War of Independence, the documentary and instrumental evidence from other parts of North America is sufficient to determine characteristic features of the various decades. The results of this study, deliberately simplified, are now considered in the light of the south-steering hypothesis.

The 1770s are indeed generally dry, not only in New England but also, by the second half of the decade, in the semiarid belt as a whole. The drought of 1777/1778, according to Hawley, almost exterminated the Hopi Indians. A tropical warm wave is dated in Central Europe 1776/1778 and no doubt extended to the United States about this date.

A minor wet phase (r-max.) commenced in the east Rockies of Canada ca. 1785, moved south to the 40th parallel ca. 1790, and to the 30th parallel ca. 1795. The 1790s were a typical easterly decade with dryness in the northwestern (southwestern Canada) parts and wetness in the southeastern parts of the tree-ring area.

The long dry period proper thus began in southwestern Canada, reached the upper Colorado by 1798, the Gila valley in 1800, the Rio Grande by 1804, and southern California by 1805. In the 1810s, wetness affected southwestern Canada and adjacent Pacific states, but the change from dry to wet over most of the area is dated fairly precisely as 1825.

The wet phase, from 1825 to 1840, which has been associated with Little Ice Age III coincides with a cold period—both in the United States and the European continent—in the 1830s, and the wetness of this phase is unique. The whole tree-ring area was affected from Canada to Mexico more completely than in the earlier wet phases (I and II): 1601 to 1620, and the 1740s. At San Francisco, documentary evidence had independently suggested to Lynch that the period 1832 to 1840 must have been wet and, on the other side of the Rockies, references to the levels of lakes in North Dakota about 1830 and to Lake Huron in the 1830s suggest that levels were very high. Tree-ring evidence points to the greatest abnormality in the upper parts of the Arkansas and Rio Grande valleys. In the former region, for instance, a well-marked r-max. must have occurred *ca. 1830*. An r-max. is clearly indicated by the rain-gauge record at Marietta, Ohio, and by the Mississippi runoff record at Vicksburg, Miss., in each case at ca. 1827. Tree-ring evidence in the Rio Grande suggests an r-max. ca. 1836, so that once again south-steering seems to have occurred.

In low latitudes of the Old World, where reliable information is now available, this period can (see above) be subdivided into two phases: a warm-dry one, then a cold-wet one. These phases also extend to the Atlantic states of the United States, and near New York state can be dated (t-max. ca. 1826, r-min. ca. 1830; t-min. ca. 1835/1839, r-max. ca. 1839/1845).

The mid-century dry phase. The "interstadial" between Little Ice Age III*a* and III*b* is marked by a short phase of remarkable warmth (a T-max. *ca. 1850*) in the Iceland-Davis Strait region of the Arctic, and a similar warm phase in the southern hemisphere is perhaps reflected in Heusser's glacier chart (in these pages). We find, as we should expect, a dry phase in the tree-ring zone of North America.

The mid-century dry phase began in southwestern Canada abruptly about 1840 and affected the whole tree-ring area (except the extreme southeast) in the course of the 1840s. The 1840s mark the culmination of the dry phase in the Pacific states and the pioneers that trekked to the west did so in a very difficult decade. The wagon tracks left in eastern Oregon in 1849 were subsequently covered by water; the water remained until the lake dried up in a later warm period in the 1920s when the tracks were recognized and photographed. In the Mississippi valley, rainfall and runoff records indicate that the r-min. occurred as early as ca. 1840 but, in California documentary evidence led Lynch (1931) to select the period 1840 to 1847 as the driest, and to note that a contemporary diarist (Belden) remembered the dry phase as 1841 to 1851 to 1852. In the 1840s it would seem that easterly winds were more frequent than usual in the United States, but not in Canada.

Table 4*

Pressure and Pressure Gradient in the Triangle Toronto–New York –St. Johns Relative to 1841 to 1845 Means

Pressure Ins. × 100		Type	Winds (New York state)	
			W	S
1836	+6	High; west type in north	−10	+1
1837	−3	Low, but dry SW	3	−5
1838	+1		6	−2
1839	+3	E in north	−1	−1
1840	+2½		2	+2
1841	−1	NE	−5	−4
1842	1	SW	−2	2
1843	0	W	4	0
1844	1		−2	0
1845	−2		6	4
1846	1		−4	−1
1847	3	S	−3	3
1848	1		1	1
1849	6	High E	−12	−4
1850	0		5	−4

* Wind data from Hough, calculated as percentages, using method as in Schove (1953*f*, 1961/1962 and unpublished). Notice that the p-min. ca. 1841 coincides with a max. in the north and west wind components, suggesting a south steering p-min. off the Atlantic coast.

Barometer and wind records are available for this period for New England and neighboring parts of North America, and they have been used in TABLE 4.

The words "great American Desert" that appeared on some early maps across the western range, covering the western parts of Nebraska, Kansas, and Texas, are consistent with travellers' tales of the mid-century dry phase that

continued on the eastern slopes of the Rockies into the 1850s. The 1850s were indeed probably westerly in type for the rainfall pattern (from ca. 1851/1865) was like that of the 1800s and the 1930s: in each of these decades the Pacific states of the west were now wet.

The moist phase 1865 to 1885. The new moist phase might be termed Little Ice Age III*b*: it coincides with ice increases after 1863 in the Davis Strait and with the T-min. in the Arctic (see above) *ca. 1875*. The wetness spread, about 1865, east of the Rockies and then affected the United States as a whole during the Civil War decades and throughout the 1870s. In the tree-ring zone an f(R)-max. is noted in southern California and the Rio Grande *ca. 1875 to 1882*;

TABLE 5
DECADAL PATTERNS IN THE UNITED STATES
(RELATIVE TO 1911/1940 NORMALS)

Decade	Pressure mid-United States	Pressure gradient	South minus north	Rainfall		Temperature		Thirty-year type
				All United States	West Central	North Central	SE	
1881–1890	−10	(−)	(E)	(>100)	(>100)	−17	−1	
1891–1900	−10	+1	(W)	98	100	−9	−5	
1901–1910	0	−3	(E)	101	105	−6	−6	Easterly
1911–1920	+10	−7	(E)	102	103	−9	−5	P-max.
1921–1930	+5	+1	(W)	100	101	−1	+2	
1931–1940	−2	+9	(W)	97	93	+9	+4	Westerly
1941–1950	−5	(−)	(E)	(>100)	(>100)	(+)	(+)	Frontal B (?)

Pressure from *World Weather Records* (1959 only) in ×100 relative to 1921/1950 and pressure gradient data from graphs given by Kincer (1940).
Rainfall data in percentages derived from data in Tannehill (1947, Tables II and IV, pp. 240 and 244).
Temperature data from *World Weather Records* kindly supplied by Callendar (°F × 10).
North Central = Marquette and Bismarck (corrected). Other stations show a similar tendency, provided "Bismarck" record published in *World Weather Records* was corrected for "city" influence at a rate of 2° F per century.
South-east = Nashville, Galveston, and Key West.
Pressure and bracketed information added since the 1950 table (in *Quarterly Journal of the Royal Meteorological Society*) was compiled.

however, the dryness spread back northward during the 1870s and 1880s reaching southwestern Canada about 1890. In low latitudes, there are three subdivisions of this phase: warm, cold, and then warm.

The dry westerly phase 1885 to 1905. The dry phase was at its height in the 1890s—the driest decade in the semiarid zone since the 1650s or the 1580s—and the dryness then extended from central Canada to Mexico. A tropical warm phase coincided with an anticyclonic phase from the north: the P-max. (Figure 3*). The f(R)-min. in southern California, the Rio Grande, and the Colorado *ca. 1888/1889*, is significantly close to the period of the P-min. in the Pacific. The decadal patterns can now be interpreted statistically according to TABLE 5.

The transition of the 1900s. The 1900s resembled the 1860s. Wet conditions that had begun earlier (1895) in southwestern Canada affected the United

* Schove, D. J. Solar Cycles and the Spectrum of Time Since 200 B.C. This monograph.

States generally, with an r-max. ca. 1904 and ca. 1910. Wetness spread south in the middle of the decade. On the Pacific coast this was the wettest decade, but the decade ca. 1910 marks the principal r-max. in the far west and, as westerly winds decreased in frequency, the rest of the tree-ring zone gradually became wet. Indeed, pressure maps suggested an easterly phase *ca. 1905* (cf. Schove, 1950*a*, p. 161).

The wet and easterly 1910s. The wetness of the 1910s stands out in the tree-ring area as a moist phase almost comparable with the 1830s. Rainfall was much above average on the southern and eastern slopes of the Rockies. Dry conditions had nevertheless reached the Jasper area of Alberta, Canada, by 1912, and the dry phase reached the western states during the decade, the Pacific coast (sheltered from the east) being drier in this decade than in the period 1886/1895. The south steering of the wet phase about 1915 is well illustrated in a map (drawn by L. F. Page in *Monthly Weather Review*, 1937, p. 16) revealing that the period 1916/1935 was to be drier in the northern half of the United States and wetter in the southern half than the preceding period 1896 to 1915. The R-max., indicated by rainfall records and tree rings alike occurred in the northeastern part of the tree-ring area *ca. 1905/1911*, but in' most of the remainder of the area *ca. 1917/1920*; in the latter areas, the date thus almost coincides with the Pacific P-max. of the major pressure oscillation.

The transition of the 1920s. Dryness in the northwestern part and wetness in the southeastern part of the semiarid zone recalls the 1840s, the 1790s, and the 1510s. The drought (associated with an r-min. on the British Columbia coast ca. 1927) was especially severe in Washington and Oregon, and tree-ring evidence indicates the ending of the wet phase by 1929 in the upper Missouri, by 1930 in the upper Colorado, and by 1933 in the Gila River basin of southern Arizona. The easterly phase was being replaced by a westerly one and, again, this change about 1930 is reflected in the rainfall map (Landsberg, 1960, p. 1524) showing the trend between the periods 1906 to 1930 and 1931 to 1955. This time it is the southern and the leeward slopes of the Rockies that become drier.

The dry westerly 1930s. The abnormal westerly phase of the 1930s was associated with dryness generally in the lee of the Rockies and with the notorious dust-bowl droughts of the plains. The great Salt Lake in Utah now shrank to a record low level and, in the course of the decade, dryness set in in Mexico. The r-min. from Toronto, Ont., Canada, to the Rockies is dated ca. 1934 and corresponds precisely to a t-max. extending from Norway to Ontario. The R-min., indicated either by rainfall records or tree rings, is dated *ca. 1919/1925* on the western side of the Canadian Rockies but, farther south, is dated ca. 1929/1935 and corresponds closely to the T-max. of the major temperature oscillation.

The wet 1940s. Information for the 1940s was not available to me when, using south steering as a guide, I wrote (Schove, 1950, p. 161): "An easterly excess had already commenced in North Canada by 1930 and spread to South Canada during the next decade. This is presumably associated with a frontal-B stage in the U. S. A. in the forties with a reversal of trends and prospects of higher rainfall in that country."

These suggestions have now been confirmed by the maps of Rodewald (1952)

of the temperature and rainfall changes between 1941 to 1950 and 1931 to 1940. The return to the rainfall pattern of the 1910s (cf. Wagner, 1929) is linked with the change to a more easterly type. Barometric pressure was generally lower.

The drought zone continued to steer south, affecting Arizona from 1934 to 1953 and, in this region, rainfall was 20 per cent less than in the 1930s. Possibly the recent long drought in Texas in the 1950s should be regarded as its final phase in the United States. However, tree-ring and rainfall changes in the 1940s indicated a new drying phase approaching from the north, and past experience reveals that south steering cannot be relied upon for long-range forecasts south of 40° N. Tree rings, in short, prove invaluable in precise climatic chronology, although they cannot yet be used to predict the future!

Conclusions

In my paper titled "Solar Cycles and the Spectrum of Time Since 200 B.C." (this monograph), auroral and European weather history was represented in diagrams and decadal indices. In the present paper, tree-ring indices established separately are used to measure the climatic oscillations affecting the weather history respectively (1) of the subarctic, (2) of North America, and (3) of low latitudes in general. As the various indices are successively refined, it is hoped to obtain a coherent picture of global pressure and temperature fluctuations over the last 2000 years. The preceding application of this wider approach to North America since 1776 shows how the several types of oscillation discussed explain the changes of climatic pattern in the instrumental period. Transatlantic temperature waves are noted that pose problems to be solved by Anglo-American cooperation. The Spectrum of Time technique implies collaboration from various disciplines and from various parts of the world, and I should be most grateful for further information and indices that can be utilized in a project that at one time seemed almost impossible.

Acknowledgments

Most of the other contributors to this volume have also helped in supplying information for Spectrum of Time studies. Acknowledgments have been given in the appropriate articles but, as far as United States evidence is concerned, I should like to thank especially, in addition to those whose names are mentioned in the text: E. Antevs, G. W. Brier, E. R. Dewey, R. W. Fairbridge, W. S. Glock, J. M. Havens, T. N. V. Karlstrom, F. P. Keen, H. H. Lamb, H. Landsberg, I. I. Schell, Waldo E. Smith, M. K. Thomas, H. C. Willett, and L. W. Wing.

*References**

Arakawa, H. 1960. Selected papers on climatic change. Meteorol. Research Inst. Tokyo. (See his article in Weather, 1957, 12.)

Callender, G. S. 1961. Temperature fluctuations and trends over the earth. Quart. J. Roy. Meteorol. Soc. **87:** 1–12.

Cross, C. I. 1961. Abstract only in U.G.I. Stockholm Abstracts.

Dewey, E. R. The 200 year cycle in the length of the sunspot cycle. J. Cycle Research. **9**(2).

* This list omits important articles by other cited contributors whose bibliographies should be consulted separately.

DOUGLASS, A. E. 1936. Climatic cycles and tree growth. Carnegie Inst. Washington Publ. **289:** 111.
KRAUS, E. B. 1956. Discussion on his papers in Quart. J. Roy. Meterol. Soc. **82:** 358.
LANDSBERG, H. E. 1960. Note on the recent climatic fluctuation in the United States. J. Geophys. Research. **65:** 1519–1525.
LYNCH, H. B. 1931. Rainfall and stream runoff in southern California since 1789. Metropolitan Water District of So. Calif., Los Angeles.
RODEWALD, M. 1952. Geografiska Annaler. Stockholm, Sweden.
RODEWALD, M. 1956. Deutsche Hydrog. Zeit. Hamburg. **9**(Heft. 4): 182–186.
SCHOVE, D. J. 1949*a*. European raininess and European temperatures, A.D. 1500–1950. Quart. J. Roy. Meteorol. Soc. **75:** 175–179; 181.
SCHOVE, D. J. 1953*e*. MSc. Thesis. London, England.
SCHOVE, D. J. 1953*f*. Proces-Verbaux des Seances de l'Assocation de Meterologie Memoires et Discussions. Paper 10: 187–193.
SCHOVE, D. J. 1954*a*. Summer temperatures and tree rings in north Scandinavia, A.D. 1461–1950. Geogr. Ann. Stockholm, Sweden. **36**(H 1–2): 40–80.
SCHOVE, D. J. 1955*b*. The sunspot cycle 649 B.C.–A.D. 2000. J. Brit. Astron. Assoc. **66**(2): 59–61.
SCHOVE, D. J. & A. W. G. LOWTHER. 1957. Tree rings and medieval archaeology. Medieval Archaeol., **1:** 78–95.
SCHOVE, D. J. 1959. Cross dating of Anglo-Saxon timbers at Old Windsor and Southampton. Medieval Archaeol. **3:** 288–290.
SCHOVE, D. J. & A. W. G. LOWTHER. 1957. Tree rings and medieval archaeology. Medieval Archaeol. **1:** 78-95.
SCHOVE, D. J. 1961/1962. Annual winds in north-west Europe since A.D. 1625–1960. Geografiska Annaler, Stockholm, Sweden.
SCHOVE, D. J. & A. FREWER. 1961. Tree rings in the Cairngorms, 1900–1956. Scottish Forestry, **15**(2): 63–71.
SCHULMAN, E. 1956. Dendroclimatic Changes in Semi-Arid America. Univ. Ariz. Press. Tucson, Ariz.
TANNEHILL, I. R. 1947. Drought. Princeton Univ. Press. Princeton, N. J.
TITOW, J. 1960. Economic History Review. Cambridge, England.
VISSER, S. W. 1959. Med. en Verh. **75**. (Kon. Ned. Met. Inst.)
WAGNER, T. 1929. Geografiska Annaler. Stockholm. **11:** 71.
YAMAMOTO, T. 1957. Long-term variations of the jet stream (in Japanese, trans. E. R. Hope. Defense Inform. Service, DRB, Canada.) Kagahu, **27:** 630–631.

POLLEN DIAGRAMS AS EVIDENCE OF LATE-GLACIAL CLIMATIC CHANGE IN SOUTHERN NEW ENGLAND*

Margaret B. Davis†

Department of Zoology, Yale University, New Haven, Conn.

Since the publication, 10 years ago, of E. S. Deevey's pollen diagrams from Aroostook County, Me.[1]—the first pollen analyses in this country in which nonarboreal pollen as well as tree pollen was systematically counted—there has been considerable interest in the United States in the study of pollen in sediments deposited contemporaneously with the final stages of retreat of Wisconsin ice. Much of the detailed work that has been done in the last few years has been devoted to the study of late-glacial sediments in southern New England. It is the purpose of this paper to review the evidence of climatic change that has been obtained in this region and to evaluate its significance.

Sediments of comparable age in northwestern Europe have been studied in great detail. In Denmark, for example, a consistent series of changes in pollen percentages and in the macrofossil and pollen flora in sediments indicate that, following the retreat of the ice, there was first an interval of tundralike vegetation (the Older Dryas period), then development of park tundra or forest (the Alleröd period, a time of more or less temperate climate), and subsequent retreat of forest as subarctic conditions returned during the Younger Dryas period.[2] In some regions there is also evidence for an older climatic oscillation, the Bölling, in sediments underlying this sequence. Radiocarbon dates indicate that these two climatic oscillations occurred between about 13,000[3] or 12,500[4,5] years ago and about 10,500[27] years ago.

Stratigraphy

Evidence for vegetational change in southern New England is subtle and difficult to interpret. The late-Glacial pollen sequence first described by E. B. Leopold from sediments at Durham, Conn., has since been found in sediments from at least 5 additional sites in Massachusetts and Connecticut[6,7,8,13] (see FIGURE 1). Radiocarbon dates from a few of these sites support their correlation with the Durham deposit.[9] In contrast to evidence from Europe, northern Maine,[1] and Nova Scotia,[11] the Durham pollen sequences seem to imply the existence of forest at the time the Two Creeks forest bed in the United States and Alleröd deposits in Europe were deposited, and its persistence during the subsequent time of climatic cooling in Europe and the advance of Valders glacial ice in the United States. The evidence for vegetational change during the Valders interval consists only of changes in the relative frequencies of some of the tree pollen types.

The pollen diagram shown in FIGURE 2 (series I) is similar to the Durham sequence. Both pollen diagrams in the figure are from a small deposit of late-glacial sediment exposed in open section at Cambridge, Mass. The vertical axes in the figure represent the height of samples in the section, above

* The work described in this article was supported in part by the National Science Foundation, Washington, D.C.

† Present address: Department of Botany, University of Michigan, Ann Arbor, Mich.

the underlying glacial outwash gravel (Lexington outwash);[12] the horizontal axes represent the percentages of each pollen type, calculated as per cent total pollen found in the samples. Tree-pollen types are shown in the left-hand portion of the diagram, shrubs in the central portion, herbs to the right, and cryptogams and aquatics to the far right. A few pollen types that occurred in extremely low frequencies are included in such categories as "other trees" and "other shrubs." A more complete discussion of the diagrams is given elsewhere.[13]

According to radiocarbon determinations from other sites, the boundary between pollen zones A-1-2-3 and A-4 (at 130 cm. in sample series I) is similar in age to the boundary between the Alleröd and Younger Dryas periods in Europe, and slightly younger than the Two Creeks forest bed in Wisconsin.[7,9]

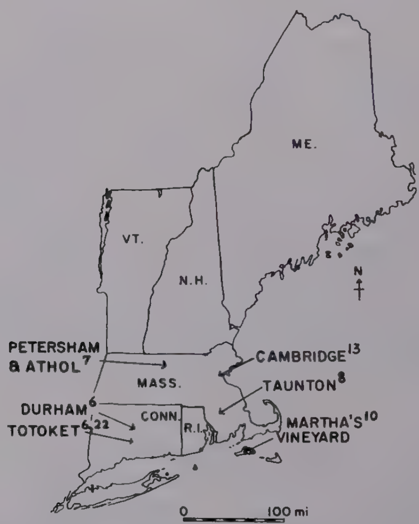


FIGURE 1. Outline map of New England, showing locations of sites discussed in text. Superscript numbers refer to references appended to the article.

We might therefore expect that the pollen diagram would record at these levels any changes in the vegetation associated with a change from interstadial to stadial climate. The pollen diagram shows only a decrease in the percentages of pine (*Pinus*) and deciduous tree-pollen types such as oak (*Quercus*), hornbeam (*Ostrya*), ash (*Fraxinus*), and beech (*Fagus*) and an increase in the percentages of spruce (*Picea*), fir (*Abies*), and larch (*Larix*) pollen. This change has been interpreted as evidence for an increase in the relative frequency of spruce in the forest, in response to a colder or more moist climate.[6,7,10,14] As is typical for the lower portion of Durham sequences, high herb-pollen percentages occur at the base of the profile (pollen zone T, below 50 cm. in series I and II). Increased percentages of tree pollen are found in the overlying sediments (zone A-1-2-3). These changes in pollen percentages have been used as evidence for the gradual development of mixed spruce and deciduous forest during an interstadial interval.[6,7,10] Radiocarbon dates from

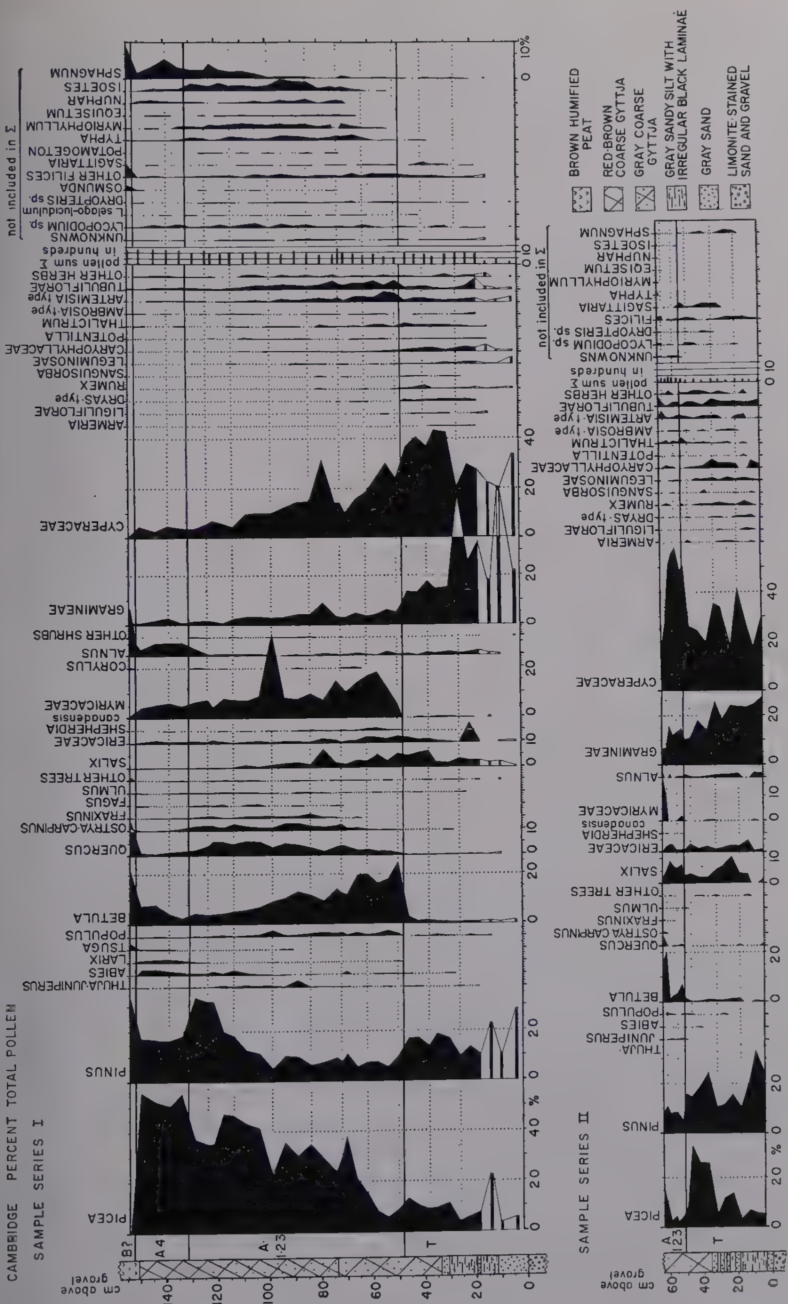


FIGURE 2. Pollen diagrams from two series of samples collected from a late-Glacial deposit exposed in open section at Cambridge, Mass. For explanation see text.

sites in southernmost New England indicate that this interval lasted at least 2000 years, from about 12,800 to 10,800 years before present (B.P.).[9,15,16]

In addition, two pollen diagrams from southernmost New England have been interpreted as evidence for climatic changes that occurred before the Durham sequences were deposited. Changes in the relative frequencies of herb and tree pollen, in sediments at the base of profiles in which a sequence similar to Durham was found, appear to indicate successive advance and retreat of forest in response to climatic changes. At Duarte's bog, Martha's Vineyard, Mass.,[10] a recent radiocarbon date from pollen zone V-2, 9910 ± 440 years B.P. (Y-646-1), fails to indicate that the sediments there are older than Durham sequences at other sites. The date is possibly somewhat too young due to contamination of the sample with younger sediment at the time of collection.[16] At Totoket, Conn.,[6] correlation of the sediments at the base of the profile with Bölling deposits in Germany (zones Ia or Ib)[3] is confirmed by a radiocarbon date from zone T-2, 13,280 ± 420 years B.P. (Y-502).[9] However, still older dates (Y-446d: 14,790 ± 160, Y-285: 13,550 ± 460) have been obtained at Totoket from the overlying sediments,[9] indicating to me that Durham and "pre-Durham" sediments at Totoket may be very close in age. Additional investigations in southern New England have failed to demonstrate pollen sequences similar to the Totoket sequence.[8,10] The Durham sequence, in contrast to the Totoket and Duarte's bog sequences, has been found repeatedly, and it seems to represent regional and contemporaneous changes in the kinds and amounts of pollen contributed to sediments.

Pollen Percentages

In the climatic interpretation of pollen diagrams from New England we have relied heavily on vegetational interpretations of pollen percentages. The assumption has been made that the percentage of a pollen type in sediments is similar to the percentage of the corresponding species in the vegetation at the time the sediments were deposited. A recent study of the present vegetation and the pollen now being deposited in lakes in northern Vermont[17] has made me less confident of this method of interpretation. The percentages of pollen now being deposited in sediments in Vermont are very different from the species percentages in the forest; some tree species are hundreds or even thousands of times better represented by pollen, in proportion to their frequency in the vegetation, than others. A small change in the vegetational frequency of a species that contributes pollen abundantly can thus cause large changes in its pollen percentage, as well as in the pollen percentages for all the other species, even though their abundance in the vegetation remains almost constant. Large changes in frequency among the minor pollen producers can probably take place without causing noticeable changes in the relative frequencies of the pollen deposited. Unless corrections are made for differences in representation, this source of error can cause difficulties both in the correlation of pollen diagrams and in their climatic interpretation.[18] The nature and magnitude of the apparent vegetational change indicated by the change in pollen percentages at the time of advance of Valders ice therefore remains problematic; a climatic interpretation would be premature. It is equally difficult to interpret changes in the relative frequencies of herb and

tree pollen in the lower portion of the Durham sequence that comprise the pollen evidence for gradual development of forest in response to climatic amelioration.

The problem of interpretation of pollen percentages is magnified where the proximity of certain plants to the site of deposition has led to their greater representation by pollen, in proportion to their abundance, than plants occurring in the "regional" vegetation. In such cases, small changes in the local vegetation, brought about by the arrival of new species, disturbances of the habitat by frost action,[19] or yearly variations in the factors favoring the pollen production of these species,[20] can affect the amounts of local pollen contributed to the sediments. This will of course affect the percentage values for the pollen contributed to the sediments by the regional vegetation.

The effect of this source of error is illustrated by the pollen diagrams from Cambridge. Sample series I and sample series II represent two vertical series of samples collected about 5 feet apart in the section. Sample series II was collected from sediments stratigraphically equivalent to sediments below the 60 cm. level in series I (sediments above 65 cm. had been removed by slumping where series II was collected). The pollen flora found at equivalent levels in the two series of samples is identical, but the percentages in which the pollen occurs are very different. These differences are too great to be attributed entirely to statistical error. Leaves and seeds from herbaceous plants occurred abundantly in these sediments,[21] indicating that herbaceous vegetation grew nearby. I feel that many of the differences in pollen percentages in the two series of samples may be due to inequalities in the amounts of herbaceous pollen contributed by the local vegetation to different parts of the deposit. At the 25-cm. level in series I, for example, there is a maximum of grass (Gramineae) pollen that does not occur at the equivalent level, about 10 cm., in the other sample series. At 45 cm. in series II, where sedge (Cyperaceae) pollen percentages are lower than in the other sample series, there is an apparent maximum of spruce (*Picea*) pollen. The overlying sediment (55-cm. level) contains high percentages of grass and sedge pollen, and correspondingly low percentages of tree pollen.

The sequence of dominant pollen types in series II, from herbs, to spruce, to herbs, with the second herb maximum "delineated at its lower and upper borders by major peaks in the *Betula* curve,"[6] is similar to the "pre-Durham" pollen sequence at Totoket, Conn.[22] However, similar changes in pollen percentages do not occur in sample series I. The lack of consistency in the pollen percentages, even within a single deposit, indicates that the pollen frequencies have been influenced by local factors and that they may be extremely misleading if used as evidence for regional vegetational changes. It follows that changes in pollen percentages found only in a single sample series from one deposit, as is the case at Totoket and Duarte's bog, Martha's Vineyard, cannot be accepted with confidence as the regional pollen sequence, of potential climatic significance, until corroborated by investigations at several sites nearby.

Presence or Absence of Pollen

Although pollen percentages may not be useful for interpretation, the presence or absence of pollen types does give some information about the vegeta-

tion. At the Cambridge site and at other sites where the Durham sequence has been demonstrated there is a gradual increase, upwards in the section, in the number of tree and shrub species and genera represented by pollen. In contrast, there is a gradual decrease upward in the number of herbaceous genera represented by pollen in the sediments. A similar change from herbaceous to arboreal species occurs in the macrofossil flora found at Cambridge.[21] The change in flora supports Leopold's[6] conclusion that the landscape became forested gradually during an interval in part contemporaneous with the Two Creeks forest, and that it remained forested at a time contemporaneous with the Valders advance. However the conclusion, based on the increase in deciduous tree pollen percentages, that temperatures rose gradually in this region, reaching a Late-Glacial maximum about 11,000 years ago,[6] is not clearly supported. In the absence of macrofossils of hardwood trees and of knowledge of the distances to which their pollen can be dispersed, it is uncertain whether these species grew in the area. The gradual increase in the number of species represented by pollen might have resulted from changing conditions of pollen dispersal, or from the migration of these species toward or into southern New England. The increase in their pollen percentages may have been caused either by increased abundance of these species within the area contributing pollen, or by a decrease in the total pollen production of local vegetation.

Macrofossils

Fortunately a few macrofossils have been found that give some indication of the late-Glacial climate. Occurrences of spruce and poplar as macrofossils at several sites, including Cambridge,[7,21,23] definitely indicate that the climate at the time pollen zone A-4 and at least a portion of zone A-1-2-3 were deposited was sufficiently warm to allow the growth of boreal species. These fossils give no indication of climatic change. However, the increase in forest, indicated by the changing pollen flora in the basal portion of the Durham sequence, does indicate climatic amelioration, if a tundra climate prevailed at the time deposition of the sequence began.

A vegetation floristically similar to tundra is indicated by arctic-alpine herbs found as macrofossils in the lower portion (zone T) of the Cambridge deposit.[21] *Salix herbacea*, *Dryas integrifolia*, and *Vaccinium uliginosum* (var. *alpinum*?) are shade-intolerant plants; they indicate that trees did not grow in the immediate vicinity of the site. These species do not occur as weeds in temperate regions, and therefore appear to require a climate colder, at least at certain times of the year, than the present climate of southern New England.[21] However, they do not imply a climate too cold for all tree growth, because they now occur in areas of discontinuous forest along the northern edge of the boreal forest, where their abundance may be controlled in part by the distribution of unshaded habitats. Such habitats occur where edaphic or microclimatic factors discourage tree growth. The presence of boreal species at Cambridge at the time these plants grew there is indicated by the occurrence of a water beetle, *Deronectes griseostriatus*, in the same sediments. This species has seldom been collected in tundra regions in North America, and never in the high arctic.[21] The pollen flora implies a similar mixture of species. Pollen

of temperate herbs and of several boreal and deciduous tree genera is present, as well as pollen of at least one arctic-alpine genus, *Armeria*. *Armeria* pollen has been found only at Cambridge and Martha's Vineyard,[10] and not at inland sites studied in comparable detail. Differences in pollen flora and pollen percentages suggest a difference between the vegetation only 20 or 30 miles inland and the vegetation along what is now the coastline;[8] it may be hazardous to generalize about the climate of New England on the basis of the Cambridge deposit. I suggest tentatively that the evidence presently available at Cambridge implies a treeless vegetation and a climate that might allow tree growth in favorable habitats.

However, we are hampered in drawing conclusions from these data by lack of knowledge of the present-day ecology of plant species, and by the small numbers of species identified. Climatic inferences drawn from the occurence of a few individuals may be unreliable because of possible evolutionary changes. Even small differences in physiology would be important from our point of view. Certainly times of changing environment, rapid changes in population size, and isolation of small populations from their parent populations would favor evolutionary changes.

Discussion

Although there is now evidence for a vegetation floristically similar to tundra in southern New England immediately following deglaciation, there is really no very conclusive evidence for a tundra climate: a climate too cold to allow the growth of trees. This is important, because it influences the climatic interpretation given to the overlying deposits. If trees were limited in their growth by climate, their gradual increase in southern New England during this interval, indicated at Cambridge and other sites where a Durham pollen sequence has been demonstrated, might be due to a gradual amelioration of climate. On the other hand, if the climate was very favorable for tree growth from the beginning, some other factor must have limited the growth of trees at first. I suggest that a factor of considerable importance may have been, very simply, the proximity of seed source.

Migration of trees is a slow process, because seeds must be transported to a favorable site, germinate, and grow for several years before the new generation can also produce seed. Migration therefore involves a number of plant generations. Its speed depends on the distance to which seed is dispersed and on the length of the life cycle of the plant, which may be influenced to a certain degree by climate. The rate of glacial retreat from New England during the period we are considering was very rapid, even when the most conservative estimate of the distance between drift borders (Middletown, Conn. to St. Johnsbury, Vt.)[24] is used as the basis for calculation. Assuming a time interval of 2000 years,[9] the rate of retreat over this distance would have averaged about 150 m./year. This rate exceeds estimated and observed rates for spruce migration in North America by a factor of 2 to 10. It is 2 times greater than the rate of post-glacial spruce migration in Sweden, and approximately equal to the rate of post-glacial spruce migration in Germany.[25] If glacial retreat was accomplished by mass wastage of ice, rather than by the orderly retreat of a continuous ice sheet, the rate at which scattered areas of bare ground suitable

for colonization by plants became available may have been still more rapid. Under these circumstances it seems doubtful that the migration of trees, at least in some parts of New England, could have kept up with the rate of glacial retreat, even if the climate had been favorable.

The possibility that factors other than climate limited the distribution and abundance of plants in newly-deglaciated areas greatly increases the difficulty of climatic interpretation of late-glacial fossil deposits. The differences between coastal sites, on the one hand, and inland sites on the other, may be related to the proximity of plant refuges, rather than to differences in climate. The transitions from herbaceous to arboreal vegetation recorded by apparently similar Durham pollen sequences may have taken place at different times in different areas. Furthermore, if trees were migrating rapidly into a region favorable to their growth, their occurrence in the vegetation and their frequency, which we have attempted to measure as an indication of climate, might not have been affected by minor climatic changes. It is therefore conceivable that several climatic oscillations are included in the interval recorded by our pollen diagrams. In this case, the "interstadial" interval, represented by pollen zones T and A-1-2-3, corresponds to more than one interstadial as recorded in other regions. The radiocarbon dates now available seem to support this hypothesis, indicating a time span of at least 2000 years, in contrast to estimates of 1000 to 1500 years for the lengths of the Alleröd period in Europe,[4,26,27] and the Port Huron-Valders interstadial in the United States.[23]

Long-distance correlations are often based on evidence for similar oscillations in climate. In correlating New England pollen deposits with glacial deposits, the possibility should be kept in mind that climatic changes that influenced the ice margin may not have affected the kinds and amounts of pollen contributed to sediments in southern New England.

Acknowledgment

I am indebted to Clifford A. Kaye, United States Geological Survey, Washington, D.C., for helpful discussion and suggestions.

Note added in proof. A pollen diagram from Red Maple Swamp in southern Connecticut,[28] although similar to the Durham sequence, shows lower percentages of nonarboreal pollen in silt at the base of the profile (zone T-2) than in the immediately overlying silt and gyttja (zone T-3). Diagrams from Tom Swamp and Pleasant St. Bog at Petersham and Athol show similar sequences within zone T-3 that are interpreted to be the result of inception of deposition of inorganic sediments at a time when there was no local vegetation and all pollen was windblown from the south.[7] In contrast, the Red Maple Swamp sequence is considered evidence for a pre-Durham vegetational and climatic change, in which the more severe interval represented by zone T-3 can be correlated with a glacial readvance in Connecticut.[28] If this interpretation is accepted, a similar interpretation and correlation should be given to the pollen sequences from central Massachusetts.

References

1. Deevey, E. S. 1951. Late-glacial and postglacial pollen diagrams from Maine. Am. J. Sci. **249:** 177–207.

2. Iversen, J. 1954. The late-glacial flora of Denmark and its relation to climate and soil. Danm. Geol. Unders. II R. **80:** 87–119.
3. Firbas, F., H. Müller & K. O. Münnich. 1955. Das wahrscheinliche Alter der späteiszeitlichen "Bölling" Klimaschwankung. Naturwissenschaften. **42:** 509.
4. de Vries, H., G. W. Barendsen & H. T. Waterbolk. 1958. Groningen radiocarbon dates II. Science. **127:** 129–137.
5. Tauber, H. 1960. Copenhagen radiocarbon dates IV. Am. J. Sci. Radiocarbon Suppl. **2:** 12–25.
6. Leopold, E. B. 1956. Two late-glacial deposits in southern Connecticut. Proc. Natl. Acad. Sci. **42:** 863–867.
7. Davis, M. B. 1958. Three pollen diagrams from central Massachusetts. Am. J. Sci. **256:** 540–570.
8. Davis, M. B. 1960. A late-glacial pollen diagram from Taunton, Massachusetts. Bull. Torrey Botany Club. **87:** 258–270.
9. Deevey, E. S. 1958. Radiocarbon-dated pollen sequences in eastern North America. Veröff. Geobot. Inst. Rübel. **34:** 30–37.
10. Ogden, J. G., III. 1959. A late-glacial pollen sequence from Martha's Vineyard, Massachusetts. Am. J. Sci. **257:** 366–381.
11. Livingstone, D. A. & B. G. R. Livingstone. 1958. Late-glacial and postglacial vegetation from Gillis Lake, Richmond County, Nova Scotia. Am. J. Sci. **256:** 341–359.
12. Judson, S. 1949. The Pleistocene stratigraphy of Boston, Massachusetts and its relation to the Boylston Street Fishweir. *In* The Boylston Street Fishweir II: 7–48. Papers Peabody Found. Arch. **4.**
13. Davis, M. B. Pollen analysis of a late-glacial deposit in Cambridge, Massachusetts (manuscript).
14. Leopold, E. B. 1958. Some aspects of late-glacial climate in eastern United States. Veroff. Geobot. Inst. Rübel. **34:** 80–85.
15. Rubin, M. & C. Alexander. 1960. U.S. Geological Survey radiocarbon dates V. Am. J. Sci. Radiocarbon Suppl. **2:** 129–185.
16. Stuiver, M., E. S. Deevey & J. L. Gralenski. 1960. Yale natural radiocarbon measurements V. Am. J. Sci. Radiocarbon Suppl. **2:** 49–61.
17. Davis, M. B. & J. C. Goodlett. 1960. Comparison of the present vegetation with pollen-spectra in surface samples from Brownington Pond, Vermont. Ecology. **41:** 346–357.
18. Fagerlind, F. 1952. The real signification of pollen diagrams. Bot. Notiser. **1952:** 185–224.
19. Sigafoos, R. S. 1951. Soil instability in tundra vegetation. Ohio J. Sci. **51:** 281–298.
20. Scamoni, A. 1955. Beobachtungen über den Pollenflug der Waldbäume in Ederswalde. Z. Forstgenetik. **4:** 113–122.
21. Argus, G. W. & M. B. Davis. Macrofossils from a late-glacial deposit at Cambridge, Massachusetts (manuscript).
22. Leopold, E. B. & R. A. Scott. 1958. Pollen and spores and their use in geology. Smithsonian Inst. Rept. **1957:** 303–323.
23. Flint, R. F. 1956. New radiocarbon dates and late-Pleistocene stratigraphy. Am. J. Sci. **254:** 265–287.
24. Flint, R. F. 1953. Probable Wisconsin substages and late Wisconsin events in northeastern United States and southeastern Canada. Geol. Soc. Am. Bull. **64:** 897–920.
25. Löve, D. 1959. The postglacial development of the flora of Manitoba: a discussion. Can. J. Botany. **37:** 547–585.
26. Nilsson, E. 1960. Södra Sverige i senglacial-tid. Geol. Fören. i Stockholm Förh. **82:** 134–149.
27. Iversen, J. 1953. Radiocarbon dating of the Alleröd period. Science. **118:** 9–11.
28. Beetham, N. & W. A. Niering. 1961. A pollen diagram from southeastern Connecticut. Am. J. Sci. **259:** 69–75.

PALYNOLOGY AND THE CLIMATIC RECORD OF THE SOUTHWEST

Paul B. Sears
Yale University, New Haven, Conn.

The southwestern United States contrasts strikingly with Scandinavia, where the statistical analysis of sedimentary pollen and other spores was first used as a clue to past vegetation and climate. In humid, glaciated northern Europe peat deposits are abundant, and they serve as an admirable preservative for these microfossils. Differences in bog strata and their organic inclusions, as well as certain facts of plant distribution, had long since aroused interest in post-Glacial climatic fluctuations. Thanks largely to pollen analysis the major fluctuations are now fairly clear and are known to correspond in general with episodes in North America.

The last major ice advance took place 16,000 to 18,000 years ago. The warm and more or less arid conditions terminating it were reversed at least twice: by the Port Huron readvance, circa 12,000 to 13,000 B.P., and by the Valders, circa 11,000 B.P., with the Two Creeks interval, circa 11,500 B.P., between them. The Valders was followed by a prolonged period of warming and desiccation, relieved by the humid Atlantic about 5000 to 6000 B.P. Cooler and moister conditions were initiated around the beginning of the Christian era, and fluctuations have occurred since that await further study.

The challenge of climatic history in arid regions far beyond the limits of continental glaciation is even greater. Old terraces and lakes, dry or saline, along with numerous evidences of human activity, historic and earlier, suggest less arid conditions in the past. However, peat is scarce or absent except at high altitudes, while lake sediments are strongly inorganic. Accordingly, for a long time the most effective clue has been the record of tree-ring patterns, cleverly extrapolated back to about the beginning of the Christian era.

This tree-ring analysis, begun by A. E. Douglass, had the double incentive of archaeological and astronomical interest, since the question of periodicity in solar behavior is an important one. However, like studies of post-Glacial peat, it deals with a limited and recent time span. On the other hand, any record that might be recovered from sedimentary basins in the Southwest would not have been interrupted by the physical presence of glaciers in these basins, and so might be expected to be more or less continuous back through the Pleistocene, or even beyond. If available, it would indicate something of the extent and intensity of Pleistocene climatic changes far outside the limits of continental glaciation, assist in correlating the numerous surface features associated with Pleistocene events in the region, and contribute to our understanding of biogeography. In addition, because of its continuity, such a record might eventually furnish data needed to investigate the principles as well as the pattern of climatic change.

The possibility of using pollen analysis in the Southwest was first evident from sediments collected by Antevs in 1935 to 1936 and studied at Oklahoma. These sediments came from temporary alluvial lakes in the Tsegi River Basin

of northeastern Arizona and contained sufficient pollen to give good evidence of climatic change. Because of initial difficulty in concentrating the pollen (later solved) and Antev's original suggestion that the beds sampled probably could not be correlated, only one of the profiles was published (Sears, 1937) and then merely to illustrate the possibilities. Meanwhile analyses by Margaret Kaeiser and further comments by Antevs have appeared (FIGURE 1). An erosion interval about 1300 A.D. and potsherds of earlier date help piece together to some degree an otherwise disjointed and fragmentary record.

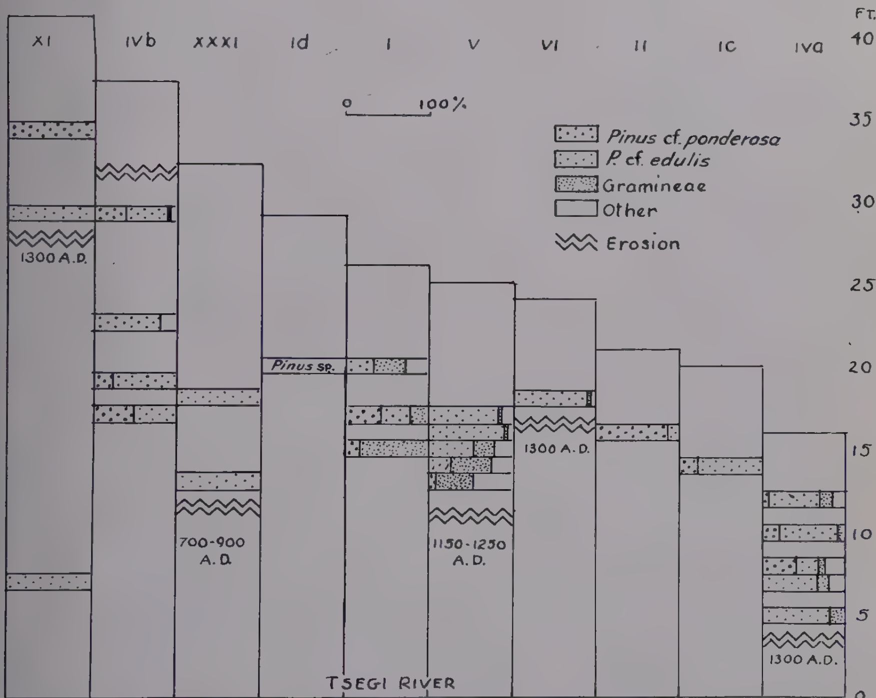


FIGURE 1. Pollen spectra from organic seams in terraces above Tsegi River, Arizona. Field numbers across top, height in feet above river at right. Collections, archeological dates, and levels of disconformity from Antevs. Analyses by M. Kaeiser.

Four vegetation zones appear to be represented in the pollen spectra. One characterized by large pine pollen (cf. *Pinus ponderosa*?) may indicate the coolest and least arid of these. A second, with smaller pine pollen (cf. *P. edulis*?) may indicate conditions now prevailing at lower, warmer, dryer altitudes. Concerning the evidence for dry grassland and still dryer desert we may speak with more assurance.

It is at least evident that the source of small pine pollen was dominant at intervals before and after 1300 A.D. and that a strong showing of larger pine pollen preceded and followed these phases, with fluctuations yet to be determined. Clearly an intensive application of palynology in the region is warranted. Since charcoal is present in some of the buried layers, radiocarbon

dating should be possible, despite the highly inorganic character of the sediments.

The general topography and climate of the Mexican plateau extends a short way into the Southwest, but as Martin *et al.* properly stress (in material unpublished) there are climatic and other differences between the Mexican plateau and the Great Basin-California type of desert. Deevey (1944) studied 6.2 m. of sediment from Lake Patzcuaro, finding pollen of pine, oak, alder, fir, grass, chenopods, composites, and tule or typha, with maize and agave as evidence of cultivation toward the top. This important pioneer paper emphasizes the advantage of lake sediments over peat for palynological study and calls attention to the need for and possibility of such study in the Southwest of the United States.

Deevey also examined archaeological material from the Basin of Mexico but found it negative, in so far as pollen was concerned. In 1948, however, with the encouragement of Pablo Martinez del Rio and the active assistance of several of his colleagues, I was able to secure a number of profiles of forest pollen from sediments in and around Mexico City, D.F., and to correlate them with cultural material (Sears, 1952). Field observation in Mexico, later in Honduras and Guatemala, shows that oak, alder, and fir indicate moister conditions than pine. On this basis there is evidence of a moist interval at depths of about 5 m. in 9 profiles at 2 sites north of Mexico City, D.F., and consistent evidence of subsequent desiccation.

A layer of volcanic ash near the top of this dry period furnished a convenient reference, as did a late, once-dry, Preclassic site, now exposed by drainage but known to have been under 2 m. of water at the time of the Spanish conquest in 1519. Except for 3 truncated profiles, all showed evidence of a return of moister conditions above the ash layer. Since the late Preclassic corresponds to the earliest Teotihuacan, our assumption is that the shift of cultural activity to higher ground ca. 500 B.C. was due to the drying up of the valley lakes, while the subsequent decay at Teotihuacan and the return of activity to the basin floor was made possible by refilling of the lakes subsequent to the beginning of the Christian era. This assumption is supported by our knowledge of the vital role of lake-margin gardens or chinampas in the Nahua economy of the Basin of Mexico.

Later, through the good offices of A. R. V. Arellano, Leonardo Zeevaert furnished samples from 2 precision cores, one of 69.05 m. in length from Bellas Artes and another of 74.9 m. in length from Madero. Pollen was analyzed by Kathryn Clisby (Clisby and Sears, 1955) and sediments by Foreman (1955). The same antithetic behavior of moist and dry indicators or, as Iversen suggested (J. Iversen, 1956 personal communication) a fluctuation of moist versus dry indicators against a background of pine continues to the base. The moist pulsations agree in a general way in both profiles and are of enough intensity to represent major Pleistocene events. Whether they prove to be in or out of phase with ice advances must await further study although it is possible (Sears and Clisby, 1955, p. 526) that ice accumulation in the North might be occasioned by increased intensity of solar radiation and consequent greater evaporation from the tropical oceans, increasing rainfall within the Basin.

Returning from Mexico, I visited the caldera known as Valle Grande in Sandoval County, N.M. and through the kindness of H. T. Stearns secured bags of chip samples taken at 10-foot intervals from a test-core 65 m. in depth. Later we obtained hand-augur samples to a depth of 14 m. Both were analyzed by Kathryn Clisby and showed a marked alternation of dry and moist conditions: actually dry-warm and moist-cool, as indicated respectively by oak and pine versus spruce and fir. The 14-m. profile is particularly good, but until dates can be obtained any attempt at interpretation would be premature.

Meanwhile Dick's report (Mangelsdorf and Smith, 1949, p. 1) on the accumulation of corn cobs during several millenia in Bat Cave, Catron County, N.M., in the cliffs of former lake St. Augustin, led us to explore the sediments in that playa. After preliminary tests by hand sampling, a 625-foot core was obtained, analyzed and published (Clisby *et al.* 1956; also 1957; and Foreman *et al.*, 1959). More recently a second core has been taken to a depth of 2000 feet and is now being studied by Clisby and Foreman at Oberlin, Ohio.

Present vegetation on the playa is arid, continental and saline, dominated by Chenopods, with grassland, juniper-pinyon, ponderosa pine, and fir in successively higher zones on the surrounding uplands (Potter, 1957). At present spruce is absent from the drainage basin, but occurs outside of it. A striking feature of the pollen profile is the spruce curve above the 200-foot level, indubitably representing parts of the Wisconsin age as evidenced by two dates: 19,000 $\pm$ 1500 at 19 feet and 27,000 $+$ 5000 or -3200 at 31 feet (*Magnolia*).

Equally striking is the reduced proportion of conifer pollen below 300 feet. From 525 to 850 feet semiarid scrub predominates, as indicated by pollen of *Artemisia*, Chenopods, and grass. The importance of *Artemisia* preceding the spruce, characteristic of at least the Late Pleistocene, is a widespread phenomenon and is discussed in the references above. In addition to the reports from the Union of Soviet Socialist Republics, Spain, and Colorado there cited, it has recently been reported from Italy by Hutchinson (1960).

While species of *Artemisia* are widespread, the genus in North America is a community dominant at more northerly latitudes and higher altitudes than the chenopods, thus indicating cooler conditions when abundant. If, during the late Pliocene and early Pleistocene the southwestern desert area, already desiccated as a consequence of mountain formation along the Pacific coast, remained at low altitude, *Artemisia* rather than *Picea* would seem to be the logical indicator of cold climatic intervals. Only later, if and when general regional uplift occurred would the invasion of spruce and the assumption of its role as a low temperature indicator become a reasonable possibility.

In the Mojave desert in California, material from a 104-foot core made available by R. F. Flint has been analyzed by Roosma (1958, FIGURE 6). Below 90 feet of salt a mud layer about 12 feet thick is dated as between 11,810 $\pm$ 140 (Y-574A) and 23,610 $\pm$ 1750 (Y-577A) B.P. On this basis it was correlated by Flint and Gale (1958) with the "classical" Wisconsin glaciation of the Great Lakes region. Samples of pollen in the salt above reflect the dominance of nonarboreal desert vegetation, while those in the mud layer show considerably more arboreal pollen, especially that of juniper, which now occurs at higher altitudes. There is a second mud layer below the first, sepa-

rated from it by salt. This has been dated by Flint and Gale as more than 29,000 years old (Y281) and preliminary examination of a sample of this material has already shown that conifer pollen is abundant.

A number of sites in New Mexico and the Texas High Plains have been investigated under the direction of Wendorf, to whom I am indebted for the following summary:

"A preliminary two-year study of the Late Pleistocene and recent environment of the Llano Estacado, the southernmost section of the Southern High Plains, has disclosed evidence for three major pluvial periods of Wisconsin age. Several minor fluctuations within each of the major intervals were recorded also. The earliest known artifacts of human origin occur only with the last of these pluvial periods. The sequence of climatic intervals was defined and dated by radiocarbon as follows:

Years B. P.	*Period*
7000–13,500	San Jon Pluvial
13,500–15,500	Monahans Interpluvial
15,500–23,000	Tahoka Pluvial
23,000–30,000	Rich Lake Interpluvial
30,000+	Early Wisconsin Pluvial

"This preliminary study, which was supported by two National Science Foundation grants, was intended as a demonstration of the feasibility of closely coordinated research by several disciplines on the ecology of a restricted area; and also as an evaluation of several previously untried techniques in radiochemistry and pollen analysis. The results of this study are now in press and will appear as a joint publication of the Fort Burgwin Research Center and the Museum of New Mexico.

"The major contribution of the project was the definition of a sequence of vegetation communities, and from this and associated invertebrate assemblages, was derived a reconstruction of changes in temperatures and precipitation from the present back for a period of approximately 30,000 years. The results are highly controversial, but if they are correct, they show that our previous conclusions about the Pleistocene environment of the High Plains was erroneous. For example, analysis of pollen from four localities indicated that extensive pine and spruce woodlands or forests covered the Southern High Plains during at least one of the pluvial periods (the Tahoka), and that at least a few such trees also were present during the other two pluvial periods. The vegetation during interpluvials was primarily grasslands, similar to today, but it was slightly cooler and more moist. Invertebrate and diatom analyses generally confirmed the vegetation reconstructions, and together with stratigraphic studies, indicated that the San Jon Pluvial consisted of at least three, and possibly four, episodes of generally wetter and cooler climate, broken by periods of reduced moisture when the temperatures were still relatively cool" (Wendorf, 1961).*

Evidence of a former expansion of pine woodland is of special interest in

* This reference was not available when the text was prepared.

view of the much earlier reports of spruce pollen in Texas (J. E. Potzger and B. C. Tharp, 1947) and Louisiana (C. A. Brown, 1938). While a restudy of the Texas deposits considerably reduces the early reports of *Picea* (Graham and Heimsch, 1960) it records 3 per cent at 13 feet in the Gause bog and estimates its disappearance at about 12,500 B.P.

Involved here is the question raised by Sharp (1953) apropos of the presence of *Acer saccharum* (sugar maple), *Carya ovata* (hickory), *Liquidambar styraciflua* (sweet gum) and *Nyssa* (*tupelo*) in northern Mexico. Are these relicts of southward migration during the Pleistocene or are they, as Martin and Harrell (1957) suggested, remnants of Tertiary distribution? The answer hinges on whether a favorable corridor existed in southern Texas for Pleistocene movement. Martin and Harrell (1957) in their able paper on biogeographical evidence, thinks not. It is well, however, to recall that a now submerged continental shelf was available for southern migration in the Gulf region; such a shelf furnished a route for northward migration along the Atlantic during the late Pleistocene (Sears, 1941). It may not have been necessary for a deciduous forest environment to extend from the Rocky Mountains to the present Texas Coast in order to permit a pathway from southeastern United States into Mexico.

Martin and his colleagues at Tucson are now engaged in pollen studies of a high order of competence, the results of which should go far toward clarifying our present knowledge when they are published.

I am deeply indebted to Martin for a manuscript copy of his joint monograph now in press. It contains a thorough account of Arizona climate and vegetation, a discussion of the limitations of pollen analysis supported by numerous experiments, but expresses confidence in the general reliability of major trends in pollen profiles. Especially important is his differentiation of the effects of summer and winter rainfall upon vegetation and human activity. In designating pollen zones he follows the practice of soil scientists in numbering from top to bottom, which seems more desirable than the reverse procedure used in geological stratigraphy, as it permits indefinite extension backward in time, an increasingly useful procedure.

Martin's present post-Glacial record begins with an initial arid period in Zone VI, followed by more humid conditions marked by increasing tree pollen in Zones V-III and a return to more arid conditions in Zones II-I representing the last 3000 years. He adduces sound reasons for believing that extinction of Pleistocene mammals was not a function of drought, presents evidence of preceramic cultivation of maize and suggests that 13th century pueblo abandonment was due to an erosion cycle plus summer rainfall deficiency rather than absolute drought. For further important details we must await his publication.

Bent (1960) has analyzed the sediments from Deadman's Lake on the crest of the Chuska Mountains of New Mexico (ca. 9000-foot elevation). If I interpret her diagrams correctly, she finds evidence of a formerly more open type of vegetation than at present, marked by abundant *Artemisia*, then after a hiatus in the record, high spruce quickly giving way to pine, the present dominant. She is properly cautious in making any interpretation, suggesting

however that the transition to spruce from a more open type of vegetation may have been a phase incident to rising temperature, as the subsequent replacement of spruce by pine must have been.

A pollen laboratory is to be set up shortly at the new Fort Burgwin research center near Taos, N.M., in connection chiefly with archaeological work. The Denver branch of the United States Geological Survey is now well staffed and equipped for pollen-and-spore analysis. Its work includes much besides the Pleistocene, but will be of increasing value in that connection as time goes on. Meanwhile I am indebted to Estella Leopold of the Survey for the following information. In sampling two deposits of the Teewinot formation near Yellowstone Park representing sediments from an Upper Pliocene lake, she found that a profile near the margin of the lake was rich in NAP while one in deeper water showed almost none, yet the ratios of AP in both profiles were the same.

Since the grassland province intervenes between the Southwest and the classical Pleistocene Northeast, the record of climatic change in this transition area deserves much more study than it has received. Lane's (1931) analysis of the McCulloch bog in northern Iowa is an excellent beginning and shows the same essential sequence, but with different indicators, as the Great Lakes and New England. Now at hand and being studied at Yale is material from Lake County, Ill., and the Sand Hill Lakes of Nebraska.

Another aspect of the Southwestern problem is the importance of sound sedimentology as a check on pollen analysis. The proportion of inorganic sediments is high, diluting the pollen rain during periods of rapid erosion. Work such as that of Foreman in Mexico and New Mexico is essential to a proper interpretation of these cores. So, too, are critical studies of contemporary pollen spectra in relation to present vegetation as shown in the beautiful studies of Davis (1960) in Vermont, Martin *et al.* (in press) in Arizona, and Potter and Rowley (1960) in New Mexico. In addition to surface samples and other pollen traps, cattle tanks are proving a useful source of information in the Southwest.

Finally, because of the great importance of Southwestern archaeology, the potential value of palynology in cave deposits should be considered. Two caves, Fishbone and Guano, above Lake Winnemucca in northwestern Nevada, demonstrate the problem. Pollen, waterlaid at the base and windborne above it, is abundant and spectra are clear. But human occupation introduces uncertainties in the way of lenticular strata and possible digging. So, too, does the recurrence of similar spectra whose relative age would be evident in a more orderly stratigraphy.

Samples of the sediments in these two caves were collected by Orr (1956) under considerable difficulty and analyzed by Sears and Roosma at Yale. The first lot consisted of spot samples from some of the several levels recognized by the collector in each cave. On our request for a more continuous stratigraphic sampling Orr was able to supply only a limited sequence from each deposit. Several attempts were made to correlate our spectra, but each time the collector, on the basis of archaeological information, pointed out errors: an illustration of the hazard of separate and remote collector and analyst.

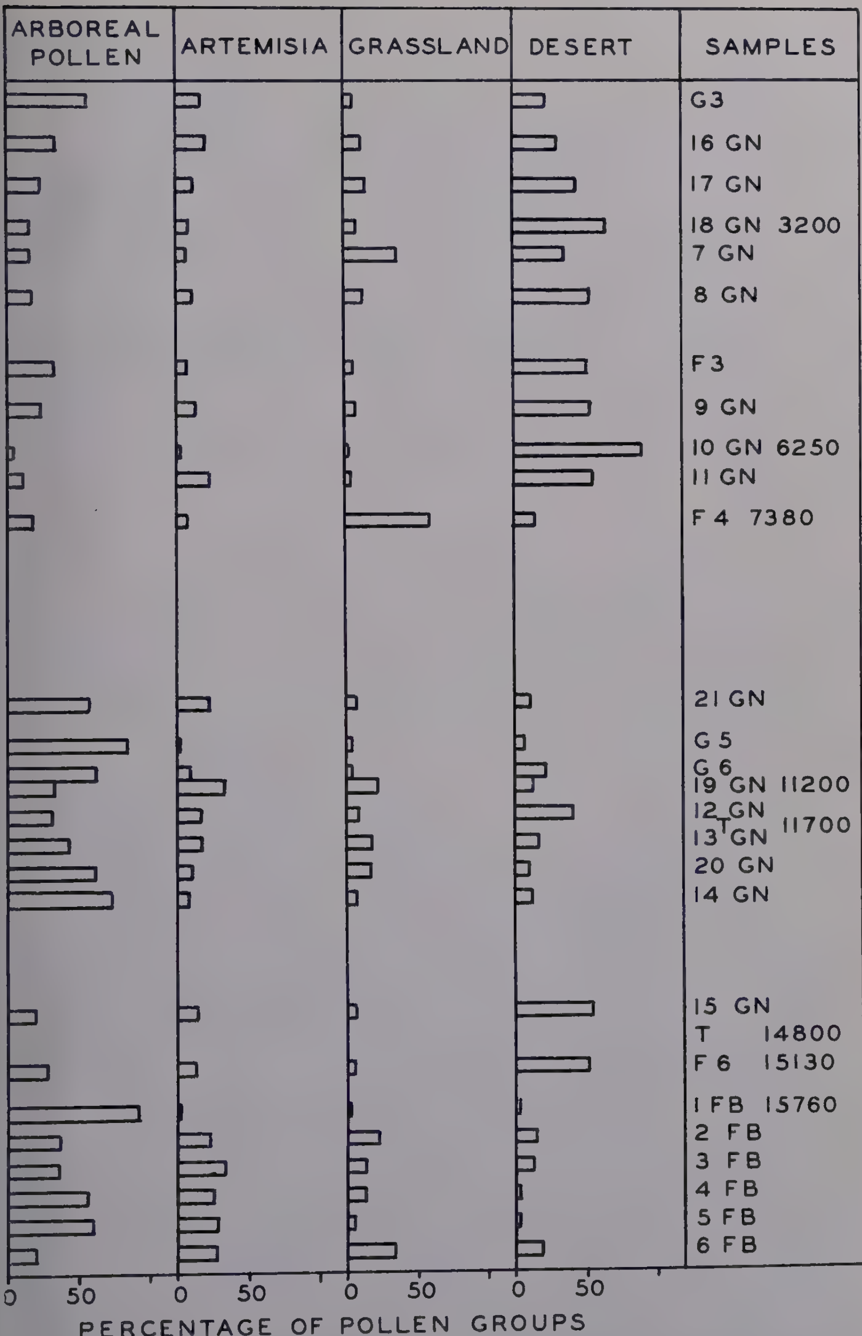


FIGURE 2. Reconstructed climatic sequence from Guano and Fishbone Caves, Nev. Collections by P. Orr. From Sears and Roosma (in press). Reproduced by permission of the *American Journal of Science.*

Finally, by supplementing field records of the sediments with our own examination of them and by grace of an accumulation of radiocarbon dates, a sequence harmonizing with known geologic, climatic, and archaeological events has been worked out (FIGURE 2). Lacking as yet is a type regional profile that could be obtained from the sediments in the lake remaining below and that should be of extreme archaeological value. However, the cave profile clearly reflects the high forest count of the late Wisconsin, the subsequent desiccation, and a moist pulsation of about 5000 years ago.

As Deevey (1944) has pointed out, there is a distinct need for pollen analysis in the Southwest. This need is now being met. Deevey further says that pollen grains must be properly preserved, which is fortunately the case, as we have shown. Again, he emphasizes the need of a standard pollen sequence or type profile for the region: now on its way in places. Finally, he stresses the importance of knowledge of the regional plant ecology. The meaning of these suggestions is clear. Wherever palynology appears to be a fruitful adjunct to other research, the specialist in that field should be brought into the project at its inception.

References

BENT, A. M. 1960. Pollen analysis of Deadman Lake, Chuska Mountains, N.M. M.S. Thesis, Univ. of Minn. pp. 1–22.

BROWN, C. A. 1938. The flora of Pleistocene deposits in the western Florida parishes, West Feliciana parish, and East Baton Rouge parish, Louisiana. *In* Contributions to the Pleistocene history of the Florida parishes of Lousiana. Geol. Bull. No. 12, Dept. of Conservation, Louisiana Geol. Survey. : 59–96.

CLISBY, K. H. & P. B. SEARS. 1955. Palynology in southern North America. III. Microfossil profiles under Mexico City correlated with the sedimentary profiles. Bull. Geol. Soc. Am. **66:** 511–520.

CLISBY, K. H. *et al.* 1956. San Augustin Plains. Science. **124:** 537–539.

CLISBY, K. H., F. FOREMAN & P. B. SEARS. 1957. Pleistocene climatic changes in New Mexico, U.S.A. Trans. Fourth Intern. Session Botan. Quaternary. *From* Geobot. Inst. Rübel. **34:** 21–26.

DAVIS, M. B. & J. C. GOODLETT. 1960. Comparison of the present vegetation with pollen-spectra in surface samples from Brownington pond, Vt. Ecology. **41:** 346–357.

DEEVEY, E. S., JR. 1944. Pollen analysis and Mexican archaeology. An attempt to apply the method. Am. Antiquity. **10:** 135–149.

FLINT, R. F. & W. A. GALE. 1958. Stratigraphy and radiocarbon dates at Searles Lake, California. Am. J. Sci. **256:** 689–714.

FOREMAN, F. 1955. Palynology in southern North America. II. Study of two cores from lake sediments of the Mexico City basin. Bull. Geol. Soc. Am. **66:** 475–509.

FOREMAN, F., K. H. CLISBY & P. B. SEARS. 1959. Plio-pleistocene sediments and climates of the San Augustin Plains, New Mexico, with a discussion by Charles E. Stearns. 10th Field Conference, N.M. Geol. Soc. : 117–120.

GRAHAM, A., & C. HEIMSCH. 1960. Pollen studies of some Texas peat deposits. Ecol. **41:** 751–763.

HUTCHINSON, G. E. 1960. History from sediment. Sci. News Letter. **78:** 343.

LANE, G. H. 1931. A preliminary pollen analysis of the East McCulloch peat bed. Ohio J. Sci. **31:** 165–171.

MANGELSDORF, P. C. & C. E. SMITH, JR. 1949. New archaeological evidence on evolution in maize. Botany Museum Leaflets, Harvard Univ. **13:** 213–247.

MARTIN, P. S. & B. E. HARRELL. 1957. The Pleistocene history of temperate biotas in Mexico and eastern United States. Ecology. **38:** 468–480.

MARTIN, P. S., J. SCHOENWETTER & B. C. AMES. (MS) Southwestern Palynology and Prehistory: the last 10,000 years. Univ. of Ariz., Tucson, Ariz. In press.

ORR, P. C. 1956. Pleistocene man in Fishbone cave, Pershing County, Nevada. Nevada State Museum Bull. No. 2, 20 pp.

POTTER, L. D. 1957. Phytosociological study of San Augustine Plains, New Mexico. Ecol. Monogr. **27:** 113–136.

POTTER, L. D. & J. ROWLEY. 1960. Pollen rain and vegetation, San Augustin Plains, New Mexico. Bot. Gaz. **122:** 1–25.

POTZGER, J. E. & B. C. THARP. 1947. Pollen profile from a Texas bog. Ecology. **28:** 274–280.

ROOSMA, A. 1958. A climatic record from Searles Lake, California. Science. **128:** 716.

SEARS, P. B. 1937. Pollen analysis as an aid in dating cultural deposits in the United States. : 61–66. *In* Early Man. G. G. MacCurdy, Ed. Lippincott. New York, N.Y.

SEARS, P. B. & K. H. CLISBY. 1955. Palynology in southern North America. IV. Pleistocene climate in Mexico. Bull. Geol. Soc. Am. **66:** 521–530.

SEARS, P. B. 1941. A submerged migration route. Science. **94:** 301.

SEARS, P. B. 1952. Palynology in southern North America. I. Archeological horizons in the Basins of Mexico. Bull. Geol. Soc. Am. **63:** 241–254.

SEARS, P. B. & K. H. CLISBY. 1952. Two long climatic records. Science. **116:** 176–178.

SEARS, P. B. & A. ROOSMA. A climatic sequence from two Nevada caves. Am. J. Science. In press.

SHARP, A. J. 1953. Notes on the flora of Mexico: world distribution of the woody diocotyledonous families and the origin of the modern vegetation. J. Ecol. **41:** 374:380.

WENDORF, F. (Ed.) 1961. Paleoecology of the Llano Estacado. Museum of N. Mex. Press, No. 1, Fort Burgwin Research Center, p. 144.

SOME COMPARISONS BETWEEN CLIMATIC CHANGES IN NORTHWESTERN NORTH AMERICA AND PATAGONIA

Calvin J. Heusser
American Geographical Society, New York, N.Y.

One of the purposes of constructing chronologies for different parts of the earth is to discover the degree of synchrony in climates and related events. The extent of synchrony has direct bearing on the solution of the problem of causes of climatic fluctuations. If it can be demonstrated that chronologies of past climates are, for the most part, synchronous for different continents in the polar hemispheres, a cause common to the earth as a whole would appear to be in effect. On the other hand if the fluctuations lack accord, an explanation would probably require more than a single basic factor. Moreover since some hypotheses set forth to explain climatic vicissitudes and the succession of glacial and interglacial ages involve out-of-phase fluctuations between the polar hemispheres, another of the purposes of chronological-environmental study is, therefore, to amass data that bear on the problem of hemispheric relationships, thus eliminating hypotheses not consonant with the accumulated facts.

The body of data gathered since the development of radiocarbon dating has placed on firmer ground the proposition that climatic fluctuations, since the wastage of the last ice age glaciers have run in phase in the Northern Hemisphere (Iversen, 1953; Lawrence, 1958; Deevey, 1958; and others). These findings have given support to the feeling that a major underlying causal source is responsible for climatic changes, and an increasing number of investigators believe this source will be found in solar variations (Lawrence, 1950; Willett, 1953; Bell, 1953; Flint, 1957; and others). However, in terms of the entire earth, available evidence favoring a single principal factor is inconclusive. Chronological data relating to climatic changes in the Southern Hemisphere are badly lacking, and connections between land masses in the hemisphere and with places north of the equator are flimsy.

The object of this paper is to discuss some recent work undertaken in Chile with reference to chronological data from related areas in South America and the Southern Hemisphere and from northwestern North America, at the same time bringing together some of the literature pertinent to late-Pleistocene* hemispheric correlation. The paucity of information from the Southern Hemisphere has long been recognized, and the American Geographical Society, New York, N.Y., realizing this, spent 2 months at Laguna de San Rafael (46° 40′ S, 74° 00′ W) in southern Chile during the 1959 winter investigating post-glacial climatic fluctuations and glacier variations (Heusser, 1960*a*). This work forms a sequel to a long-term research program the society began some years ago to study late-Pleistocene climates and climate-glacier relationships on the North American North Pacific coast.†

* Late Pleistocene, as used here, corresponds to the late-Glacial and post-Glacial, or any part thereof, as treated by pollen stratigraphers. This usage is in keeping with the belief that the Pleistocene epoch has not come to an end.

† Most of the research undertaken by the society on the North Pacific coast and in Chile.

The Problem of Correlation

Before discussing the relations between the hemispheres, some difficulties involved in interpreting chronological data in terms of climatic changes should be noted. Seemingly anomalous behavior is oftentimes encountered that can be placed in its proper perspective only through cognizance of certain relevant factors.

The extent of synchrony to be expected in order to establish an in-phase relationship is, of course, variable. It is illogical to expect perfect harmony of fluctuations because of the many situations that control the opening and close of a given event at different places. For example, a radiocarbon date for a sample taken from the base of a peat deposit and implying a time of glacier recession can be spurious with regard to that event, if biotic migration is not taken into account and the fact that the earliest organic matter was not deposited until after biota first began thriving at the site. Thus since ice sheets, within limits, receded latitudinally following the last glaciation, dates of samples from progressively higher latitudes can be younger than the recession that occurred at these places because of the migration lag. This can also be true for the dates of samples from increasingly higher elevations in mountainous areas. Moreover, dates for glacier recession from oceanic compared with continental areas might show a wide range of variation. In considering a delay, however, environmental influences should be kept in mind, because differences can be caused by local climatic conditions and need not be the result of tardy migrations of biota.

Along this line of reasoning, attention should be given to the time span between the beginning of a climatic fluctuation and the creation of a glacier that advances and inters material that is sampled, then used to date the advance and infer fluctuation. On the upper slopes and valley headwalls in mountainous country or in relatively high latitudes where glaciers originate, glaciation will be recorded earlier than in the lowlands bordering mountains or at places in low latitudes. This general situation points to the likely reason for disharmony between times for a climatic change dated from a glacier advance and dated from a bog deposit that has been zoned by means of peat and pollen stratigraphy.

Certain regions can be influenced only locally by climatic change and, in this way, anomalies arise. Short-term variations that take place in upland areas because of the influence of an altitudinal factor are likely to have no recognizable counterpart at low elevations at points distantly removed from, or even in proximity to, the upland. A small change may in fact occur, but fossil biota do not reveal it. This circumstance develops because the magnitude of the change is not great enough to exceed thresholds or tolerance limits of even critical biota, and thereby disturb the biological equilibrium effected in the region by the set of broad climatic conditions prevailing.

Local crustal movement can affect limited glaciations or glacier advances that have few or no correlatives, and no alteration of climate need be attendant. Regions where mountain building is known to have occurred and is continuing

was supported by contract with the Office of Naval Research, Washington, D.C., as arranged by the Geography Branch.

at the present time are suspect. This mechanism involves the following modifications: (1) an increase in the area above the regional snow line or the firn limits of glaciers; (2) the moving of the mountain mass upward into a zone of maximum snowfall; and (3) where crustal movement occurs in sudden shocks, pronounced avalanching and accumulation at the heads of glaciers.

Brief mention should be made of some other anomalous situations. One concerns the unusual variations of small glaciers whereby, without an apparent alteration of climate, the snout suddenly advances until the accumulation basin is discharged and then retreats, the process being repeated when sufficient ice has again accumulated to bring about advance. Another takes into account the extremely rapid advance of a large glacier during a year or less, at a time when regional ice bodies are dwindling, and a maximum is attained strikingly out of proportion with maxima of neighboring glaciers. Still another case involves certain glaciers (very few, however) that for many decades continue to move forward, completely out of phase with the regional trend. However, glaciers are not the only examples. Recurrence horizons, which generally result from climatic change causing drying and humification at the surfaces of muskegs and bogs, can vary in number, position, and extent in the deposits because of local changes in drainage conditions and water-table relations. Some variation can be effected by the differential regression of marine waters during post-Glacial isostatic crustal adjustment.

In addition to the foregoing examples that contribute to the problem of correlation, errors connected with the radiocarbon-dating method and with the reliability of the samples collected for dating should be recognized. These difficulties, however, have been discussed in many publications (Johnson, 1959) and need not be enumerated at this time.

The problem of correlation is obviously complicated, not only regionally but also hemispherically, particularly by the anomalies that have been mentioned above, and by the insufficiency of supporting data. Discrepancies ranging from decades to millenniums are apparent, and the juxtaposition of events must frequently rest with individual interpretation. It is plain that these features detract from making assured connections, chiefly where distantly removed points in the polar hemispheres are concerned. In the following sections, attention will be called to any apparent anomalies, and many correlations, of necessity, must be provisional, pending the collection of additional field information and critical radiocarbon dates.

Climatic Changes in Northwestern North America

The earliest late-Glacial age from the Pacific Northwest is established from samples taken from sediments 40 feet below the bottom of Lake Washington in the Puget Lowland, Wash. The age is 14,000 ± 900 B. P. (before present) and is thought to provide a minimum time for the recession of the Vashon glacial lobe in the Seattle, Wash., area (L-330, Broecker and Kulp, 1957; see FIGURE 1). Northwestward, at higher latitudes along the Pacific coast, the earliest late-Glacial dates are considerably younger: 10,850 ± 800 B. P. on Langara Island, Queen Charlotte Islands, British Columbia, Canada (L-297C, Broecker and Kulp, 1957); 10,300 ± 400 B. P. (L-297 D, Broecker and Kulp, 1957) and 10,300 ± 600 B. P. (L-297A, Olson and Broecker, 1959)

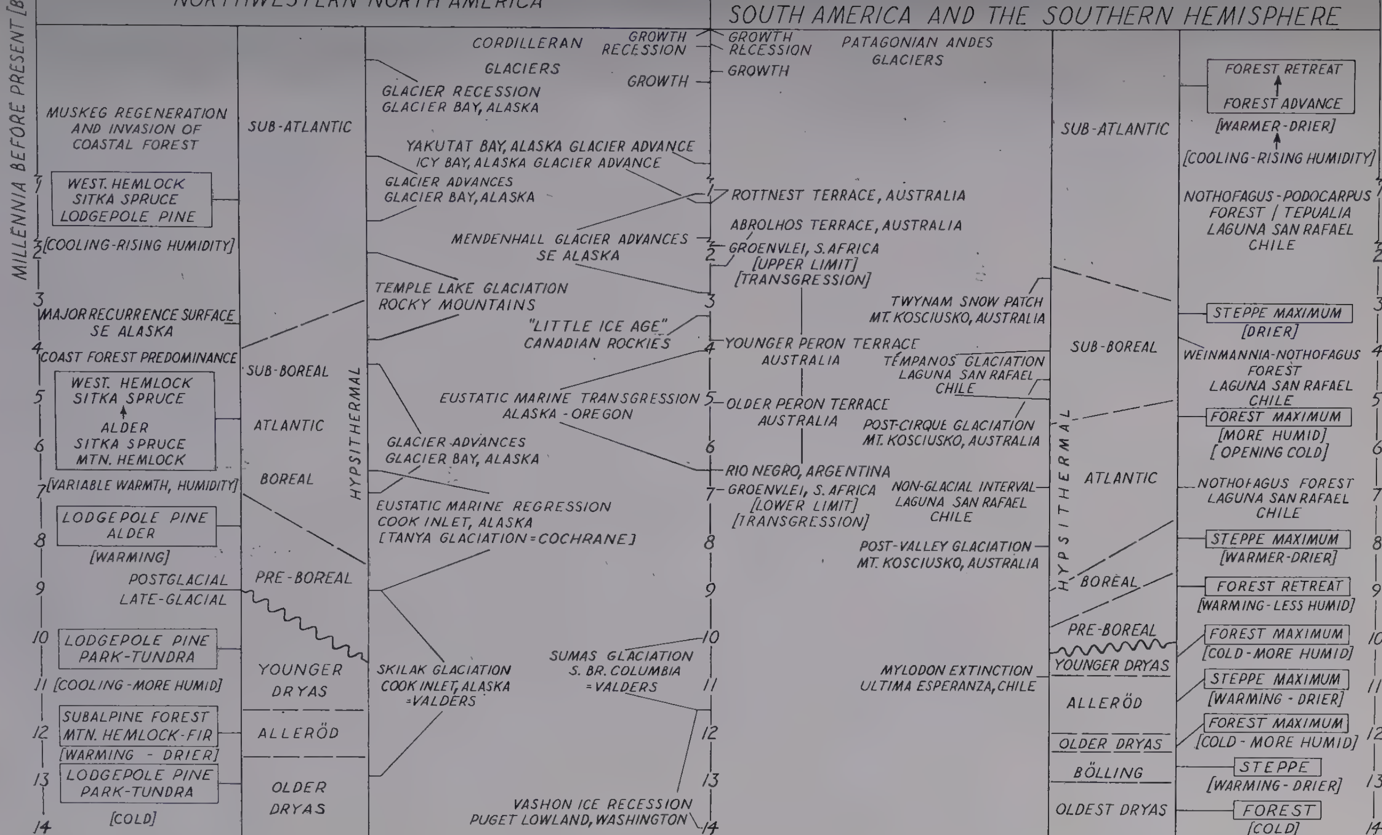


FIGURE 1. Late-Glacial and post-Glacial correlation diagram for northwestern North America and Patagonia and related areas in South America and the Southern Hemisphere. Pollen-stratigraphic zones run vertically in the middle part of each half of the diagram; climatic and vegetation successions are shown on the flanks, and on the inside of each half are some noteworthy glacial and sea-level changes. Note that the time scale for the first and second millenniums is not uniform with other millenniums. See text for further explanation.

at Juneau, Alaska; 10,820 ± 420 B. P. at Icy Cape, Alaska (I[AGS]-9, Heusser, 1959); 10,390 ± 350 B. P. at Alaganik near the mouth of Copper River, Alaska (I[AGS]-5, Heusser, 1959); and 9600 ± 650 B. P. near Ninilchik on the Kenai Peninsula, Alaska (Broecker *et al.*, 1956). These radiocarbon ages postdate Sumas glaciation in the Fraser River valley of southwestern British Columbia, which Armstrong (1956) estimates to be 11,300 ± 300 B. P. Relevant determinations are as follows: 11,450 ± 150, 11,700 ± 150, and 10,950 ± 200, respectively L-331 A, B, and C (Broecker and Kulp, 1957); and 11,500 ± 1100 and 11,000 ± 900, L-221 D and E (Broecker *et al.*, 1956). Sumas glaciation, an apparent correlative of the Valders of mid-continental North America (Broecker and Kulp, 1957) and of the Skilak of Cook Inlet, Alaska (Karlstrom, 1957), is believed to have been represented only by valley glaciers in this part of British Columbia. Northwestward along the coast, ice sheet proportions seem to have been manifest during this time. Wastage of the glacier mass did not begin until later, as indicated by the ages and distribution of radiocarbon-dated samples. Skilak glaciation of the Susitna-Cook Inlet Lowland, Alaska, is dated between 12,500 and 9000 B. P. by Karlstrom (1957, 1960) based on the samples and ages W-416: 12,900 ± 300 and W-474: 10,370 ± 350 (Rubin and Alexander, 1958) and L-137L: 9600 ± 650, L-137C: 9500 ± 650, and L-163D: 9200 ± 600 (Broecker *et al.*, 1956). It is worthwhile to point out the apparent effect of latitude in the case of Sumas and Skilak glaciations. Skilak glaciation near 60° N began about a millennium earlier and ended close to a millennium later than Sumas glaciation near 50° N.

A late-Glacial climatic record, partitioned by zones considered equivalent to the Older Dryas, Alleröd, and Younger Dryas of the European zonation, is believed to be shown by pollen profiles and peat stratigraphy from a site near Humptulips on the Olympic Peninsula of western Washington (Heusser, 1960*b*). This locality lies outside the limits of Vashon glaciation found in the Puget Lowland and along the Strait of Juan de Fuca between Canada and Washington, but was affected by proglacial disturbances. A lodgepole pine park-tundra, provisionally dated between 14,000 and 12,500 B. P., prevails during a cold interval (Older Dryas) following Vashon glaciation; mountain hemlock-fir forest, indicating ameliorating climate (Alleröd), succeeds ca. 12,500 B. P., and a subsequent return of lodgepole pine park-tundra ca. 11,500 B. P. is contemporaneous with Sumas glacier advance in the Fraser valley, Canada. Northwestward, in British Columbia and Alaska, the youngest late-glacial zone is the only one evident. The boundary between the late-glacial and post-glacial is set at ca. 10,500 B. P. in Washington and southern British Columbia, ca. 10,000 B. P. in northern British Columbia and southeastern Alaska, and ca. 9000 B. P. in south-central and southwestern Alaska.

During the early part of postglacial time, Karlstrom (1956) finds evidence in the coastal bog stratigraphy of south-central Alaska for episodes of eustatic marine regression, which he places between 9000 and 6500 B. P. while climate was cool but warming. He equates the sea-level oscillations with Tanya glaciation, his equivalent of the Cochrane readvance in eastern Canada. This event runs just prior to an interval of apparent marine transgression, dated 6500 to 4000 B. P. between southeastern Alaska and Oregon (Heusser, 1959, 1960*b*). Five coastal lakes near present-day sea level contain nonpollen de-

posits and quantities of ditch grass and goosefoot pollen indicative of tidal environments. The sequence of regression and transgression on the North Pacific coast finds general accord with the eustatic sea-level changes plotted by Fairbridge (1958, 1960).

The regressive state of sea level from 9000 to 6500 B. P. and the following transgression between 6500 and 4000 B. P. coincide with the cool but ameliorating climate of the pre-Boreal and the variable warmth and moisture of the Hypsithermal. Pollen profiles from the pre-Boreal reveal that lodgepole pine and alder were the principal constituents of the vegetation between Washington and southeastern Alaska. Northwestward, coastal tundra was manifest. A date of 7800 ± 300 B. P. (L-297G, Olson and Broecker, 1959) from a muskeg near Juneau, Alaska, and another of 8140 ± 390 (I[AGS]-12, Heusser, 1959) from a muskeg near Lituya Bay in addition to pollen control have been used as guides to set the close of the pre-Boreal between ca. 8500 B. P. in Washington and 7000 B. P. in south-central and southwestern Alaska.

The pollen-stratigraphic zones—Boreal, Atlantic, and sub-Boreal—that cover the length of the Hypsithermal, have not been recognized with certainty in northwestern North America thus far. Peat stratigraphy of certain muskegs, however, suggests a parallel zonation where two pronounced ligneous horizons (recurrence surfaces), which may represent the warmer and drier Boreal and sub-Boreal, are separated by sedge peat, equivalent to the humid Atlantic. However, this stratigraphy appears in only a few sections and is unreliable. The problem of these ligneous horizons in relation to rising and falling humidity in coastal muskegs is complex but could probably be solved by assigning radiocarbon dates to an array of horizons. One horizon, which is representative of the most pronounced recurrence surface usually encountered in muskegs, is dated 3500 ± 250 B. P. (L-106B; Kulp *et al.*, 1951) from a wood sample in a muskeg near Juneau, Alaska (Heusser, 1953). This date and two others (I[AGS]-1 and I[AGS]-15, respectively 3470 ± 180 and 2950 ± 150, from Karluk, Alaska, and Seaview, Washington; Heusser, 1959), because of their associated peat and pollen stratigraphy, have served to mark the close of the Hypsithermal and the opening of the sub-Atlantic between ca. 3000 B. P. in Washington and 4000 B. P. in south-central and southwestern Alaska.

Predominance of coast forest over muskeg is achieved during the Hypsithermal. Western hemlock, mixed with Sitka spruce in southeastern Alaska and northern coastal British Columbia and with Douglas fir in southern British Columbia and Washington, is the principal tree in the forests of the Alexander Archipelago, Alaska, and mainland south to Washington. Alder is prominent in the vegetation northwest of the archipelago, being mixed with western hemlock and Sitka spruce about Icy Cape, with mountain hemlock in Prince William Sound and with birch, which had migrated from interior Alaska, on the Kenai Peninsula and Kodiak Island.

Because climates that prevailed during the Hypsithermal were theoretically warmer than at present, glaciation during this interval is unexpected. Yet, in Glacier Bay, Alaska, at least two ice advances ca. 7000 and 4300 B. P. have been discovered (Y-10: 7050 ± 240; Preston *et al.*, 1955, and Y-302: 4330 ± 80; Barendsen *et al.*, 1957). These advances may be anomalous, the result of upward crustal movement in their source areas in the Saint Elias Mountains

Province, in Canada and Alaska, which is a notorious, tectonically unstable region. As already pointed out, the elevating of a mountain mass increases the areas and amounts of accumulation, thereby providing an increased supply of nourishment for glaciers. On the other hand, advances can also represent the response to the increase of atmospheric moisture during Atlantic time, possibly even coupled with tectonic changes. Some cooling, however, seems necessary at these times in order that the effect of summer ablation of the glacier snouts near sea level be diminished. It is also reasonable to suspect that these advances are related to the more humid climates that gave rise in muskegs to sedge peats lying beneath the two ligneous horizons discussed previously. In Cook Inlet, Alaska, Tanya and Tustumena glaciations (Karlstrom, 1960) are apparent correlatives of the Glacier Bay advances.

The sub-Atlantic is shown by pollen and peat stratigraphy to be a zone representing cooler and more moist climate. Muskeg regeneration and its displacement of forest on relatively low-lying terrain occurred at this time. Poorly humified sphagnous peat and the increasing proportions of lodgepole pine, a shade-intolerant tree relegated to open muskeg because of its inability to compete with the trees of the coastal forest, are indicative of regeneration.

Evidence for glacier advance is more abundant in the sub-Atlantic than in the Hypsithermal because of a generally cool, moist climate that was more advantageous for glaciation. Advances are recorded in Pacific coastal Alaska in Glacier Bay at ca. 1500 and 800 B. P. (Y-4: 1540 ± 130; Y-6: 1520 ± 140; and Y-7: 760 ± 130; Preston *et al.*, 1955); at Mendenhall Glacier near Juneau ca. 2800, 1800, and 1100 B. P. (Y-132-80: 2790 ± 130; L-106C: 1790 ± 285; and Y-132-84: 1090 ± 60; Kulp *et al.*, 1951; Preston *et al.*, 1955); at Icy and Yakutat Bays ca. 1200 and 800 B. P., respectively (W-374: 1200 ± 160 and W-559: 830 ± 160; Rubin and Alexander, 1958, 1960); and at Cook Inlet ca. 3000, 2000, and 1000 B. P. (Tustumena and Tunnel glaciations, Karlstrom, 1960). In the continental interior in the La Sal Mountains, Utah, Temple Lake glaciation is estimated between 3800 and 2000 B. P. (Richmond, 1960) from the radiocarbon date W-143: 2800 ± 200 (Rubin and Suess, 1955). A correlative of this glaciation is indicated by two dates (606: 3261 ± 250 and 607: 3327 ± 320; Libby, 1951) for buried peat in the Canadian Rockies near Waterton, Alta., Canada. The peat is overlain by gravels that are related to a "little ice age" cirque moraine (Horberg and Robie, 1955).

It is probable that at least some of the glacier variations on the coast are correlative, although certain local anomalous behavior resulting chiefly from crustal changes can be expected, as Miller (1958) believes. For example, on the western flank of the Fairweather Range (a part of the Saint Elias Mountains Province) the land has risen at one locality close to 100 m. relative to sea level in the last 8000 years and ca. 50 m. in the last 3000 years at another (Heusser, 1960*b*). These measurements are indicative of the magnitude of crustal movement, which could have been the instrumental factor underlying the advances discovered in Glacier, Yakutat, and Icy Bays, contiguous to the Saint Elias source region.

The glacier activity of recent centuries in the northwestern cordillera is illustrated by a generalized curve depicting advance and retreat since ca. 1600 A. D. (FIGURE 2). It has been prepared from a number of tree-ring and re-

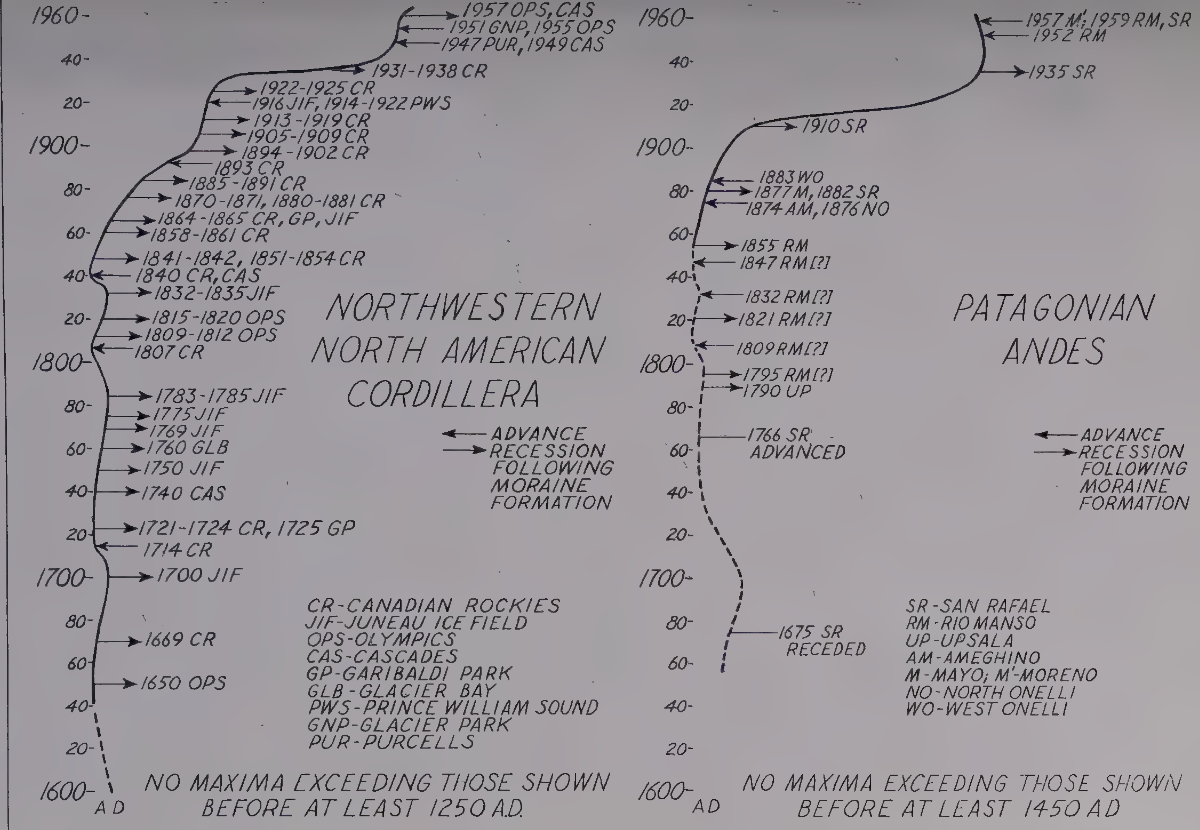


FIGURE 2. Glacier variations during the last several centuries in the cordillera of northwestern North America and in the Patagonian Andes. Curves shown depict the relative degree of advance and recession and are generalized, based on data obtained from glaciers in areas listed in the lower part of the diagram; broken lines are drawn from data of uncertain reliability. See text for further explanation.

lated studies, which also disclose that previous comparable activity has not taken place for between 500 and 700 years (Field, 1937; Cooper, 1937, 1942; Lawrence, 1948, 1950, 1953; Mathews, 1951; Heusser, 1956, 1957; and Heusser and Marcus, 1960). The salient features of this curve are chronologically: (1) a maximum about the middle of the 1600s, lasting until ca. 1700 at some places, followed by recession; (2) another maximum early in the 1700s from which retreat took place at different times up until ca. 1785; (3) readvances during the first half of the 1800s that, in a number of areas exceeded maxima of the 1700s; (4) gradual recession until the late 1800s when the rate increased up to the beginning of the 1900s; (5) slow retreat and some advances during the interval lasting into the late 1920s, after which recession occurred at an incomparable rate for the time since the 1600s; and finally (6) a decrease in the rate of retreat during the 1940s followed by thickening of glacier snouts and by advance during the 1950s. This most recent display of activity was observed in the glaciers of the Cascades and Olympics of Washington (Hubley, 1956; Bengtson, 1956; and Heusser 1957), the Purcell Mountains of British Columbia (West, 1955), and the Rocky Mountains of Glacier National Park (Dightman and Beatty, 1952), among several other areas. It had by 1960 subsided for the most part (LaChapelle, 1960), yet some glaciers in the northern Cascades and in southern British Columbia continue to creep forward (Austin S. Post, personal communication).

Climatic Changes in Patagonia and Related Areas in South America and the Southern Hemisphere

The late-Glacial of Patagonia, indeed of the Southern Hemisphere, is little known, and what information is available has largely been derived by inference and long-range correlation. Most of what is known has been provided by Auer (1958), but the chronology is generally based on parallel changes in Europe, except where there is some radiocarbon chronological control. In the accompanying FIGURE 1, the boundary for the Alleröd and Younger Dryas zones is inferred from the dates set by Godwin and Willis (1959) for the British Isles (Alleröd, 12,000 to 10,800 B. P.; Younger Dryas, 10,800 to 10,300 B. P.). The lower boundary for the Bölling, also inferred, is suggested by a date from Germany (H88-74: 13,250 ± 280, Münnich, 1957) while the upper boundary, indicating a short duration for the Older Dryas, is from a determination made on a Netherlands sample (K-542: 12,070 ± 140, Tauber, 1960).

Some significance may be attached to a date (484: 10,832 ± 400, Arnold and Libby, 1951) determined for droppings of the giant sloth (*Mylodon*) from a cave at Ultima Esperanza, Chile (Bird, 1951). The pollen content determined by Salmi (1955) revealed a complete absence of arboreal grains, which led to the conclusion that no forests or individual trees but rather steppe made up the vegetation in the surroundings. Auer (1960) believes that this vegetation and the dry prevailing climate belong to the Alleröd.

Auer's (1958) pollen and related studies form the ground work for the interpretation of late-glacial climates. His profiles show an interplay between forest and steppe vegetation whereby the relatively cool and humid forest climate fluctuates with the warmer and drier climate of the steppe. On this

foundation, Auer delimits the Bölling and Alleröd zones, times of comparative dryness and amelioration. The Oldest and Older Dryas and the Younger Dryas, which bound the Bölling and Alleröd, as shown in FIGURE 1, are colder and more humid zones when forest replaced steppe.

The boundary between the late-Glacial and the post-Glacial is conjectured from a European source (Godwin and Willis, 1959) and from a date (K-561: 8960 ± 400, Tauber, 1960) for the pre-Boreal in a deposit at Sabana de Bogotá, Colombia (van der Hammen and Gonzalez, 1960). It is between 10,300 and 10,100 B. P. The pre-Boreal, resting above the boundary but below the Boreal, closes also from the standpoint of conjectured dates between 9750 and 8600 B. P. The maximum date is supplied by Godwin *et al.* (1957) from Great Britain and the minimum by van der Hammen and Gonzalez (1960) from a Sabana de Bogotá sample zoned in the Boreal (K-560: 8020 ± 120, Tauber, 1960). Pre-Boreal climate was cold but undergoing warming and drying as manifested by the pollen record for a retreating forest.

The Hypsithermal, beginning with the Boreal zone (9750 to 8600 B. P.), extends through to the close of the sub-Boreal that is figured between 3000 and 2200 B. P. Dates (W-783: 2950 ± 200 and W-785: 2700 ± 200, Rubin and Alexander, 1960) from Páramo de Palacio, Colombia (van der Hammen and Gonzalez, 1960), and a date (Y-183-IV: 2240 ± 60; Deevey *et al.*, 1959) from Estancia Pirinaica, Tierra del Fuego (Auer, 1960), provide the time for the termination of the Hypsithermal, the sub-Boreal and sub-Atlantic boundary. The Atlantic zone, interposed between the Boreal and sub-Boreal, endured chronologically from ca. 7500 to 5000 B. P. or ca. 9000 to 5500. The time brackets 7500 to 5000 B. P. are estimates made by van der Hammen and Gonzalez (1960), and those 9000 to 5500 B. P. are from age determinations made by Deevey *et al.*, (1959) for samples collected by Auer from Tierra del Fuego at Altos Boqueron (Y-188: 8905 ± 110 and Y-189: 9380 ± 90), at Estancia Pirinaica (Y-183-II: 6600 ± 90 and Y-183-III: 4480 ± 50), and at Rio Ewan (Y-182: 6690 ± 100). It should be noted that the dates 9000, 5500, and 2200 B. P. are ages for 3 volcanic eruptions, respectively eruptions I, II, and III, which Auer (1958, 1960) utilizes for correlating his peat sections and pollen profiles.

The validity of using the Hypsithermal (an episode of greater warmth than the present) to cover the Boreal, Atlantic, and sub-Boreal in Patagonia may be questioned. Auer (1958) states that the Atlantic, or middle zone, was at first cold and moist and later warmer and drier. This conclusion stems from his pollen profiles that show forest advance at the outset and retreat at the close. The Boreal and sub-Boreal are, however, generally coincident with warmth and dryness (steppe), although early sub-Boreal time exhibited fluctuations between forest and steppe climates. It is apparent that a quantitative basis for temperature measurement is needed in order to establish the duration of the Hypsithermal in southern South America.

The moist character of Atlantic climate in Patagonia is probably correlated with "Period II" of the pollen diagrams drawn for New Zealand by Cranwell and von Post (1936) and for Hawaii by Selling (1948). Temperature conditions for New Zealand during this interval are uncertain, but because of the

latitudinal similarity to Patagonia, a cool climate is likely. Climate in Hawaii at this time was warm as well as wet, and the relative proximity of the Hawaiian Islands to the equator makes this reasonable. Deevey's (1955) study of moa remains in Pyramid Valley, New Zealand, suggests sub-Boreal drought about 3600 B. P., which is in agreement with the climate and age of this zone in Patagonia.

The findings resulting from the American Geographical Society's observations and collections at Laguna de San Rafael in southern Chile cover the interval beginning with the Atlantic zone and are in general accord with Auer's climatic sequence from this time to the present (Heusser, 1960*a*). A date (Y-737: 6850 ± 200, Stuiver *et al.*, 1960) for wood in peat underlying till of San Rafael Glacier, which in 1959 stood in proximity to the site, coincides with a time the glacier receded from the edge of the laguna into its valley that heads in the North Patagonian Icefield. A pollen diagram of the section in which the wood is intercalated indicates that the date falls within the Atlantic, and that a wet and probably cool climate became warmer and drier, as suggested by the succession from *Nothofagus* (beech) forest to forest predominated by *Weinmannia*. Where the boundary between the sub-Boreal and sub-Atlantic lies is not known. A pollen diagram from the bottom of a lake located on the morainal rim (Témpanos glaciation) that almost encircles Laguna de San Rafael reveals that *Weinmannia* continued to be an important constituent of the forest for some time after 3700 B. P. which is the date (Y-738-2: 3740 ± 400, Stuiver, *et al.*, 1960) assigned for a sample very near the base of the section.

Because peat deposits in two lakes sampled on the moraine are not thick and because a date for one gives its age at 3700 B. P., the age of the moraine and of Témpanos glaciation that it represents is figured to date from about 4000 B. P. From the age of the wood (6800 B. P.) and its position in the peat that is overlain by till near the 1959 terminus of San Rafael Glacier, the beginning of Témpanos glaciation is placed at about 5000 B. P. (Heusser, 1960*a*). Muller (1959, 1960) is, however, unable to reach accord with this dating primarily because the single till unit overlying the peat deposit near the present terminus of the glacier appears to be recent. It is possible that the glacier, although it has come forward and receded since the time till buried the near-glacier section, did not recede far enough to expose the site before the 1900s. For this reason the till appears unweathered and recent. In one respect, the 4000 to 5000 B. P. age of Témpanos glaciation is difficult to reconcile, mainly because no glacial conditions compatible with the size and extent of the Témpanos moraines are suggested in the pollen record. A piece of evidence, however, is offered from Auer (1958), who writes that after eruption II (dated 5500 B. P. or early sub-Boreal), "violent fluctuation" of steppe and forest took place. Climatic changes are indicated for ca. 5000 B. P., the postulated time of advance of Témpanos ice, when the hiatus between the laguna sections occurs. Despite these considerations, supplementary data are necessary before the age of Témpanos glaciation can be fixed.

Of interest in this connection is a series of dates recently published for samples from the Mount Kosciusko district in Australia (Rubin and Alexander,

1960). One sample, postdating valley glaciation, is 8100 B. P. (W-769: 8100 ± 250); another, postdating cirque glaciation, to which Témpanos glaciation may correspond, is 4600 B. P. (W-770: 4580 ± 220); and another dates the age of the Twynam snow patch, related to the "little ice age," at 2500 B. P. (W-768: 2520 ± 200).

Deevey *et al.* (1959) have determined a number of ages for samples connoting changes of sea level during Atlantic, sub-Boreal, and sub-Atlantic time. At Rio Negro, Argentina, two episodes of transgression date from ca. 6700 B. P. (Y-196: 7120 ± 70, Y-197: 6570 ± 110, and Y-198: 6555 ± 130). At Groenvlei, South Africa, corresponding transgression begins ca. 6900 B. P. and is followed by regression until 1900 (Y-466: 6870 ± 160 and Y-467: 1905 ± 60). In Australia, on Rottnest Island and at Point Peron, dates for transgressions resulting in the Younger Peron Terrace and the Older Peron Terrace are given, respectively, at 3800 and 5100 B. P. (Y-337: 3810 ± 90 and Y-324: 5120 ± 130).

The sub-Atlantic opened 3000 to 2200 B. P. when the comparatively mild and dry climate of the sub-Boreal changed, becoming cool and humid. The forest advanced farthest eastward into the steppe at this time, and Tierra del Fuego in southern Patagonia became almost completely forest-covered (Auer, 1958, 1960). Later, however, the forest receded, and its present position is the result of climatic fluctuation and the destructive activity of man and grazing animals. In western Patagonia at Laguna de San Rafael, coolness and increasing moisture are also evident from the pollen profiles for the section on the Témpanos moraine. The tree *Podocarpus* in association with *Nothofagus* reached increasingly greater proportions during this interval, and on poorly-drained ground *Tepualia* simultaneously increased in area.

Glacier activity in response to the generally cool, moist climate of the sub-Atlantic may have taken place in Patagonia prior to the formation of the modern moraines, but no advances have been discovered that date from an earlier part of this interval. A curve depicting the variations since the late 1600s for the glaciers studied is shown in FIGURE 2. Glaciers taken into account are San Rafael in Chile and Rio Manso in Argentina (Lawrence and Lawrence, 1959), Ameghino in Argentina (Nichols and Miller, 1951; Mercer, 1960*a* and *b*), and Upsala, Mayo, North Onelli, West Onelli, and Moreno, also in Argentina (Mercer 1960*a* and *b*). Since San Rafael Glacier has received greater attention than any of the others, it plays a more important part in the construction of the curve. No maxima greater than those shown have been achieved probably for several centuries; in the case of San Rafael, this is known to be true since at least 1450 A. D.

The Patagonian glacier curve is far less satisfactory than the one drawn for the glaciers of northwestern North America. Nevertheless, the points brought out are: (1) during the late 1600s a state of recession is indicated, but by the latter half of the 1700s advance had taken place; (2) maxima were reached in at least one case, from which recession is dated ca. 1790; (3) up until ca. 1850, evidence for the variations is found solely in the growth rings of a tree in the vicinity of the Rio Manso Glacier terminus, which depicts retreat ca. 1795, advance ca. 1809, recession, and advance ca. 1847; (4) slow re-

cession after 1855 with periodic halts and advances until ca. 1910; (5) rapid recession between 1910 and ca. 1940; and (6) thereafter little change with advances registered by at least three glaciers during the 1950s.

Regarding Polar Hemisphere Correlations

Because of the scanty chronological data available from the Southern Hemisphere, it would be untimely to reach any conclusions regarding the relationships between the polar hemispheres for all of the Late-glacial and Postglacial. The evidence from northwestern North America and from Patagonia and related regions, however, shows a general accord, compatible with the belief that climatic events are in phase. The best evidence in support of in-phase relations stems from the harmonious glacier variations since at least the middle of the 19th century.

A broad parallelism of pollen zones is seen between the North Pacific and Patagonia. It should be noted that, in the former region, some evidence for an equivalent Bölling oscillation during the late-glacial, although not entered in FIGURE 1, is found in the pollen stratigraphy. Supporting data are needed, however, in order to substantiate this claim. For the North Pacific postglacial, no pollen zones have been fixed in the Hypsithermal. Zonation is suggested at certain sites, but its widespread application does not seem feasible. In this respect, Auer's (1958, 1960) adherence to a zonal pattern in Patagonia for the time covered by the Hypsithermal seems somewhat forced. This point is illustrated by Auer's Atlantic (warm-humid zone of Blytt and Sernander) which, at some places, was at first cold and moist and toward the close became warmer and drier. Another questionable point is his sub-Boreal, when marked climatic fluctuations were manifest, and the warm, dry continental climate of the classic sub-Boreal obtained over only part of this zone.

Some synchronous glacier advance and recession is in evidence. Tanya glaciation of Cook Inlet, Alaska, shows correspondence with Mount Kosciusko, Australia, valley glaciation. Témpanos glaciation at Laguna de San Rafael, Chile, shows contemporaneity with Mount Kosciusko cirque glaciation and with ice advance in Glacier Bay, Alaska. If Témpanos glaciation, as placed in FIGURE 1, is anachronous as mentioned earlier, it may fall during the time of valley glaciation in Australia, that is, prior to the dated nonglacial interval (ca. 6800 B. P.) at Laguna de San Rafael. The age of Témpanos glaciation as indicated (5000 to 4000 B. P.) is, however, in accord with eustatic sea-level changes related to glaciation. It occurs at the time of the Bahama Emergence when sea level had dropped, and also between the times of high level corresponding to the Older and Younger Peron Terraces (Fairbridge, 1958).

The record of eustatic sea-level fluctuations on the North Pacific coast of North America is, by and large, in agreement with records collected from Southern Hemisphere coasts. The three episodes of low sea level in Cook Inlet, Alaska (9000 to 6500 B. P.) and the succeeding transgression between Alaska and Oregon (6500 to 4000 B. P.) are matched with a regressive ocean in South Africa and South America (before 7000 to 6500 B. P.) followed by intervals of transgression on these continents and in Australia (7000 to 1000 B. P.). Transgression in Australia (Fairbridge, 1958) includes the Abrolhos Terrace (ca. 2300 B. P.) and the Rottnest Terrace (ca. 1200 to 1000 B. P.),

as well as those mentioned previously. Transgression on the North Pacific coast, as far as evidence shows, occurs no later than ca. 4000 B. P., or the time of the Younger Peron Terrace (3800 B. P.).

Except for the glacier advances of recent centuries, the multiplicity of glacier variations during the sub-Atlantic in northwestern North America finds little in the way of counterparts in the Southern Hemisphere. Variations seem certain to have taken place, but the relatively few field studies made thus far have failed to uncover them. The recently determined age of the Twynam snow patch (ca. 2500 B. P.) in the Mount Kosciusko area, Australia, is significant in this regard because it fixes the time of snow line lowering and the outset of environmental conditions, which were conducive to glacier alimentation. The agreement between the age of the snow patch and glaciation in the Rocky Mountains (ca. 3300 B. P.) and Alaska Coast Mountains (ca. 2900 B.P.) is good.

To conclude, one cannot overemphasize the need for supplementary data in connection with the problem of polar hemispheric relations. It should also be pointed out that although this discussion concerns the Late Pleistocene, this need is equally great for all of the Pleistocene. Because certain parts of the time scale are better stocked with data, stress should not be placed in these areas, but rather in areas where lacunae make correlations at present extremely risky. Future chronological-environmental work should bear particularly on: (1) establishing the validity for the application of the European post-Glacial (Blytt-Sernander) and late-Glacial zonal systems between the hemispheres; and (2) ascertaining the manifold glacier variations, expressly during the late-Glacial and sub-Atlantic in the Southern Hemisphere, and eliminating from consideration all irrelevant anomalous activity.

References

Armstrong, J. E. 1956. Mankato drift in the lower Fraser valley of British Columbia, Canada. Bull. Geol. Soc. Am. **67:** 1666–1667.

Arnold, J. R. & W. F. Libby. 1951. Radiocarbon dates. Science. **113:** 111–120.

Auer, V. 1958. The Pleistocene of Fuego-Patagonia. Part II: The history of the flora and vegetation. Ann. Acad. Scient. Fennicae, III. Geologica-Geographica, **50.**

Auer, V. 1960. The Quaternary history of Fuego-Patagonia. Proc. Roy. Soc. London (B) **152:** 507–516.

Barendsen, G. W., E. S. Deevey & L. J. Gralenski. 1957. Yale natural radiocarbon measurements III. Science. **126:** 908–919.

Bell, B. 1953. Solar variation as an explanation of climate change. *In* Climatic Change. : 123–136. Harvard Univ. Press. Cambridge, Mass.

Bengtson, K. 1956. Activity of the Coleman Glacier, Mt. Baker, Washington, U. S. A. J. Glaciol. **2:** 708–713.

Bird, J. 1951. South American radiocarbon dates. *In* Radiocarbon dating. Mem. Soc. Am. Archeol. **8:** 37–49.

Broecker, W. S. & J. L. Kulp. 1957. Lamont natural radiocarbon measurements IV. Science, **126:** 1324–1334.

Broecker, W. S., J. L. Kulp & C. S. Tucek. 1956. Lamont natural radiocarbon measurements III. Science. **124:** 154–165.

Cooper, W. S. 1937. The problem of Glacier Bay, Alaska: a study of glacier variations. Geog. Rev. **27:** 37–62.

Cooper, W. S. 1942. Vegetation of the Prince William Sound region, Alaska, with a brief excursion into post-Pleistocene climatic history. Ecol. Monog. **12:** 1–22.

Cranwell, L. M. & L. von Post. 1936. Post-Pleistocene pollen diagrams from the southern hemisphere. Geog. Ann. **18:** 308–347.

Deevey, E. S. 1955. Paleolimnology of the Upper Swamp Deposit, Pyramid Valley, New Zealand. Records Canterbury Museum. **6:** 291–344.

DEEVEY, E. S. 1958. Radiocarbon dated pollen sequences in eastern North America. Veröff. Geobotanisches Institut Rübel in Zürich. **34:** 30–37.

DEEVEY, E. S., L. J. GRALENSKI & V. HOFFREN. 1959. Yale natural radiocarbon measurements IV. Am. J. Sci. Radiocarbon Suppl. **1:** 144–172.

DIGHTMAN, R. A. &. M. E. BEATTY. 1952. Recent Montana glacier and climate trends. Monthly Weather Rev. **80:** 77–81.

FAIRBRIDGE, R. W. 1958. Dating the latest movements of the Quaternary sea level. Trans. N. Y. Acad. Sci. **20:** 471–482.

FAIRBRIDGE, R. W. 1960. The changing level of the sea. Scientific Am. **202:** 70–79.

FIELD, W. O. 1937. Observations on Alaskan coastal glaciers in 1935. Geog. Rev. **27:** 63–81.

FLINT, R. F. 1957. Glacial and Pleistocene Geology. Wiley. New York, N.Y.

GODWIN, H., D. WALKER & E. H. WILLIS. 1957. Radiocarbon dating and postglacial vegetational history: Scaleby Moss. Proc. Roy. Soc. London (B) **147:** 352–366.

GODWIN, H. & E. H. WILLIS. 1959. Radiocarbon dating of the late-glacial period in Britain. Proc. Roy. Soc. London (B) **150:** 199–215.

VAN DER HAMMEN, T. & E. GONZALEZ. 1960. Upper Pleistocene and Holocene climate and vegetation of the "Sabana de Bogotá" (Colombia, South America). Leidse Geologische Mededelingen **25:** 261–315.

HEUSSER, C. J. 1953. Radiocarbon dating of the thermal maximum in Southeastern Alaska. Ecology. **34:** 637–640.

HEUSSER, C. J. 1956. Postglacial environments in the Canadian Rocky Mountains. Ecol. Monog. **26:** 263–302.

HEUSSER, C. J. 1957. Variations of Blue, Hoh, and White Glaciers during recent centuries. Arctic **10:** 139–150.

HEUSSER, C. J. 1959. Radiocarbon dates of peats from North Pacific North America. Am. J. Sci. Radiocarbon Suppl. **1:** 29–34.

HEUSSER, C. J. 1960*a*. Late-Pleistocene environments of the Laguna de San Rafael area, Chile. Geog. Rev. **50:** 555–577.

HEUSSER, C. J. 1960*b*. Late-Pleistocene environments of North Pacific North America. Am. Geog. Soc. Spec. Publ. **35**.

HEUSSER, C. J. & M. G. MARCUS. 1960. Glaciological and related studies of Lemon Creek Glacier, Alaska. Am. Geog. Soc. Final Rept. Juneau Ice Field Research Project (mimeo.).

HORBERG, L. & R. A. ROBIE. 1955. Postglacial volcanic ash in the Rocky Mountain piedmont, Montana and Alberta. Bull. Geol. Soc. Am. **66:** 949–956.

HUBLEY, R. C. 1956. Glaciers of the Washington Cascade and Olympic Mountains; their present activity and its relation to local climatic trends. J. Glaciol. **2:** 669–674.

IVERSON, J. 1953. Radiocarbon dating of the Alleröd period. Science. **118:** 4–6.

JOHNSON, F. 1959. A bibliography of radiocarbon dating. Am. J. Sci. Radiocarbon Suppl. **1:** 199–214.

KARLSTROM, T. N. V. 1956. The problem of the Cochrane in Late Pleistocene chronology. U. S. Geol. Survey Bull. **1021-J**.

KARLSTROM, T. N. V. 1957. Tentative correlation of Alaskan glacial sequences, 1956. Science. **125:** 73–74.

KARLSTROM, T. N. V. 1960. The Cook Inlet, Alaska, glacial record and Quaternary classification. U. S. Geol. Survey Prof. Paper **400-B:** 330–332.

KULP, J. L., H. W. FEELY & L. E. TRYON. 1951. Lamont natural radiocarbon measurements I. Science. **114:** 565–568.

LACHAPELLE, E. R. 1960. Recent glacier variations in western Washington (abst.) J. Geophys. Research. **65:** 2505.

LAWRENCE, D. B. 1948. Mt. Hood's latest eruption and glacier advances. Mazama. **30:** 22–29.

LAWRENCE, D. B. 1950. Glacier fluctuation for six centuries in southeastern Alaska and its relation to solar activity. Geog. Rev. **40:** 191–223.

LAWRENCE, D. B. 1953. Recession of the past two centuries. In: Periodicity of deglaciation in North America since the late-Wisconsin maximum. Geog. Ann. **35:** 83–94.

LAWRENCE, D. B. 1958. Glaciers and vegetation in southeastern Alaska. Am. Scientist. **46:** 89–122.

LAWRENCE, D. B. & E. G. LAWRENCE. 1959. Recent glacier variations in southern South America. Am. Geog. Soc. Southern Chile Expedition 1959. Tech. Rept. **1.** (mimeo).

LIBBY, W. F. 1951. Radiocarbon dates, II. Science. **114:** 291–296.

MATHEWS, W. H. 1951. Historic and prehistoric fluctuations of alpine glaciers in the Mount Garibaldi map-area. J. Geol. **59:** 357–380.

MERCER, J. H. 1960*a*. Glacier fluctuations on the eastern side of the south Patagonian Andes. Unpublished data.

MERCER, J. H. 1960*b*. Glacier fluctuations on the eastern side of the south Patagonian Andes (abst.). Bull. Geol. Soc. Am. **71:** 2104.

MILLER, D. J. 1958. Anomalous glacial history of the northeastern Gulf of Alaska region (abst.). Bull. Geol. Soc. Am. **69:** 1613–1614.

MULLER, E. H. 1959. Glacial geology of the Laguna San Rafael area. Am. Geog. Soc. Southern Chile Expedition 1959. Tech. Rept. **2.** (mimeo.).

MULLER, E. H. 1960. Glacial chronology of the Laguna San Rafael area, southern Chile (abst.). Bull. Geol. Am. **71:** 2106.

MÜNNICH, K. O. 1957. Heidelberg natural radiocarbon measurements I. Science. **126:** 194–199.

NICHOLS, R. L. & M. M. MILLER. 1951. Glacial geology of Ameghino Valley, Lago Argentino, Patagonia. Geog. Rev. **41:** 274–294.

OLSON, E. A. & W. S. BROECKER. 1959. Lamont natural radiocarbon measurements V. Am. J. Sci. Radiocarbon Suppl. **1:** 1–28.

PRESTON, R. S., E. PERSON & E. S. DEEVEY. 1955. Yale natural radiocarbon measurements II. Science. **122:** 954–960.

RICHMOND, G. M. 1960. Glaciation of the east slope of Rocky Mountain National Park, Colorado. Bull. Geol. Soc. Am. **71:** 1371–1381.

RUBIN, M. & C. ALEXANDER. 1958. U. S. Geological Survey radiocarbon dates IV. Science. **127:** 1476–1487.

RUBIN, M. &. C. ALEXANDER. 1960. U. S. Geological Survey radiocarbon dates V. Am. J. Sci. Radiocarbon Suppl. **2:** 129–185.

RUBIN, M. &. H. E. SUESS. 1955. U. S. Geological Survey radiocarbon dates II. Science. **121:** 481–488.

SALMI, M. 1955. Additional information on the findings in the Mylodon cave at Ultima Esperanza. Acta Geographica. **14:** 314–333.

SELLING, O. H. 1948. Studies in Hawaiian pollen statistics. Part III, On the late-Quaternary history of the Hawaiian vegetation. Bernice P. Bishop Museum Spec. Publ. **39**.

STUIVER, M., E. S. DEEVEY & L. J. GRALENSKI. 1960. Yale natural radiocarbon measurements V. Am. J. Sci. Radiocarbon Suppl. **2:** 49–61.

TAUBER, H. 1960. Copenhagen radiocarbon dates IV. Am. J. Sci. Radiocarbon Suppl. **2:** 12–25.

WEST, R. 1955. The recent history of the Commander Glacier. Canadian Alpine J. **38:** 99–101.

WILLETT, H. C. 1953. Atmospheric and oceanic circulation as factors in glacial-interglacial changes of climate. *In* Climatic Change. : 51–71. Harvard Univ. Press. Cambridge, Mass.

NOTES ON LATE-QUATERNARY CLIMATIC CHANGES IN CANADA*

J. Terasmae

Geological Survey of Canada, Ottawa, Ont., Canada

Introduction

In recent years research projects in the earth and biological sciences undertaken in all parts of Canada have significantly widened the range and improved our knowledge of geologic and biological events of late-Quaternary age. The scope of many of these studies has included climatic changes as a framework on which to base the interpretation of results. Advance and retreat of continental glaciers, sea-level fluctuations, migrations of plants and animals, formation of soil profiles and drainage patterns, changes in sedimentary sequence and sequences of fossils all depend in some measure on changes in climate. Knowing the present-day relationships, past climatic environments can be reconstructed with some degree of certainty.

In other parts of the world enough detailed studies have been made to enable compilation of fairly accurate climate curves and maps showing, for example, the composition of "fossil forests" at certain times in the post-Glacial episode. Although similar patterns are beginning to appear, much more careful work is required before such compilations can be made in Canada, except for local areas.

It is with this sobering thought in mind that I wish to emphasize the word notes, in the title of this article. Nevertheless, there is a reasonably large number of these notes available, and some lines of approach to paleoclimatic problems of the late-Quaternary time in Canada will be outlined in the following.

Studies pertaining to Quaternary research in Canada have passed through an encouraging development in the past 10 to 15 years, and it is the opinion of the writer, based on numerous discussions with his colleagues in various fields of science, that this has been a phase of basic studies and reconnaissance characterized by conservatism in interpretation and critical observation and gathering of field data. Only recently have many preliminary studies been completed and reports published on the obtained results.

Scientists have at times ventured guesses at paleoclimates as inferred from their studies, but the bravest lot of gamblers, by far, have been the palynologists who have had definite ideas on paleoclimates for a long time. Perhaps fortuitously such paleoclimates usually have been described in general terms, such as "cool, moist" and "warm, dry," as compared with the present climate in the region where the studies were made. Such designations of climate may be quite adequate for some purposes, but they are insufficient and even misleading for more specific discussions. For example, warm, cool, dry, and moist are relative terms, and absolute figures, even approximate, are much more descriptive and definite and can be compared with available meteorological data.

References to the literature are intended as examples only, and I make no

* This article is published with the permission of the director of the Geological Survey of Canada.

claims for adequate coverage of available published reports. Several of the references listed, however, contain useful lists on bibliography in their specific fields of study.

Sources of Information

Meteorologic observations. Records on observations of weather in Canada are available for more than 100 years in only rare cases; generally reliable meteorologic observations extend back 30 to 50 years at a number of localities. Meteorologic stations in more outlying areas are rather recent. Scattered notes on weather can be found in written records much farther back in time, but these are of less interest and value because of their discontinuous nature (Koeppe, 1931).

Geologically speaking and also from the paleoclimatologic point of view, the written records on climate in Canada are recent. Nevertheless, *such records form the only sound basis for interpretation of past climates* (Canada Year Book, 1960; Chapman, 1952; Chapman, 1938, 1953; Hare, 1950, 1952*a* and *b*; Kendrew and Currie, 1955; Kendrew and Kerr, 1955; Rae, 1951; Shenfeld and Slater, 1960; Thomas, 1953). Paleoclimatologic studies would certainly benefit if meteorologic records from more localities in Canada were available. It is hoped that the excellent work carried on by the Meteorological Branch, Air Services, of the Canada Department of Transport can be extended considerably in the future.

Glacial geology and related studies. Geological studies of Quaternary deposits made across southern Canada have established in some detail the stratigraphy and correlation of glacial, interglacial, and interstadial deposits (Antevs, 1925, 1931; Armstrong, 1957; Chapman and Putnam, 1951; Coleman, 1933, 1941; Craig, 1956; Deane, 1950; Dreimanis, 1960*a* and *b*; Elson, 1955, 1957, 1959, 1960; Fyles, 1956; Gadd, 1960; Glacial Map of Canada, 1958; Gravenor, 1957; Henderson, 1959*a*, *b*, and *c*; Hughes, 1956; Johnston, 1917; Karlstrom, 1956; Karrow, 1957, 1959; Kupsch, 1960; Laverdiere and Courtemanche, 1959; Lee, 1955, 1957; MacClintock and Terasmae, 1960; Mirynech, 1961; Prest, 1957; Stalker, 1960; Terasmae and Hughes, 1960*a*; Watt, 1953). Owing to the fact that the regime of glaciers is ultimately dependent on climate, the studies in glacial geology have provided much valuable information of paleoclimatologic importance.

Geologic studies have shown that the beginning of the last, Wisconsin, glacial stage consisted of several advances and retreats of the glacier similar in magnitude to the final retreats and advances of the Wisconsin ice. Thus it can be inferred from geologic studies that climatic oscillations leading to the Wisconsin glaciation began with a trend towards lower annual mean temperatures in southeastern Canada. Average total precipitation may have remained unchanged, but the annual snowfall in relation to rain must have increased. Lower annual mean temperatures could have been caused by lower summer temperatures; higher winter precipitation, particularly as snow, could have resulted from generally milder winters with fewer extremely cold periods. Such conditions would increase the total snowfall and reduce the melting period during summer. Further evidence for this statement will be presented later.

The main Wisconsin ice advance covered southern Canada, and thus no records of climatic conditions were preserved during the height of glaciation. However, retreat of the Wisconsin ice, interrupted by halts and readvances, left abundant evidence of climatic conditions, only a part of which has been studied. Rapid retreat of the ice margin about 11,000 to 7000 years ago is clearly indicated by abundant evidence of melt-water erosion and deposition. The magnitude of such phenomena require large volumes of water and often torrential currents. Formation of large ice-dammed lakes such as Glacial lakes Agassiz, Barlow-Ojibway and Iroquois, and particularly the great volumes of sediments deposited in them required correspondingly huge sources of supply that could be derived only from the melting of the continental glaciers.

The steep gradient of tilting of the beaches formed around the shores of some of the ice-dammed lakes indicate rapid differential uplift and release from the load of the ice. All these data indicate that at the close of the Wisconsin glaciation a significant and quite abrupt increase in temperature is required for an almost "catastrophical" melting of the continental glaciers.

Geologic studies made in southern Canada stimulated the extension of similar studies northward through the provinces and into the Northwest Territories (Antevs, 1925; Bird, 1959; Cook, 1960; Craig, 1960; Craig and Fyles, 1960; Geol. Survey of Canada Bull. 54, 1959; Glacial Map of Canada, 1958; Hare, 1959; Henderson, 1959*b*; Hughes, 1956; Ignatius, 1956; Ives, 1959*a*, *b*, 1960; Laveriere and Courtemanche, 1959; Lee, 1959, 1960; MacKay, 1958; Nicholson, 1960; Prest, 1957; Terasmae and Hughes, 1960*b*). Much important paleoclimatological information has been gleaned from these investigations. First, the last remnants of ice melted from areas east and west of Hudson Bay about 7000 to 6000 years ago. It has been postulated that annual mean temperatures then may have been about 5° F. warmer than at present (Ives, 1959*a*). However, following the hypsithermal interval in central Labrador, a cooling of climate with resulting greater accumulation of snow is indicated by formation of the nivation hollows (Henderson, 1959*b*). Further evidence for such a cooling of climate is found in advances of alpine glaciers in western Canada and Alaska (Goldthwait, 1960; Karlstrom, 1960). The history of the Tyrrell Sea (the post-Glacial marine inundation in the Hudson Bay region) has been outlined by Lee (1960), and this study is also pertinent to climatic interpretations in that region.

Studies of permafrost, its regime, and southern limit have proved to have paleoclimatologic implications (Cook, 1960).

Certain biologic, geographic and archaeological studies in the Canadian Arctic have in several ways contributed to climatologic considerations (Bird, 1959; MacKay, 1958; Meldgaard, 1960; Nicholson, 1960; Porsild, 1955, 1957, 1958; Terasmae and Craig, 1958; Terasmae, 1959).

Studies of modern glaciers in the Arctic and the Cordillera have yielded climatologic information extending from the present to some considerable antiquity (Hattersley-Smith, 1958, 1959; Mathews, 1951).

Oceanographic studies. Evidence of former higher and lower sea levels has been recognized for many years, but only recently have such phenomena been related more closely to changes in climate (Fairbridge, 1958, 1960; Shepard and

Suess, 1956). Along the coasts of Canada, however, studies of the past sea-level fluctuations have been complicated by isostatic and other crustal movements that have been of varying magnitude in different regions.

Much more useful paleoclimatological information has been gleaned from studies of fossils in marine sediments (Wagner, 1959) coupled with radiocarbon or other methods of dating of marine fossils (Elson, 1960). Stratigraphic, paleontologic and sedimentologic studies of Quaternary marine sediments in Canada, now in progress, are already yielding promising data also on paleoclimatic problems. Such studies in the Arctic will probably eliminate much speculation inherent in current theories on the Ice Ages.

Paleontologic studies. Investigations of Quaternary vertebrate fossils (Hibbard, 1949) in Canada have been made in some areas widely scattered in both space and time (for earlier studies see Baker, 1920), and there is much need for more paleontologic studies. Taxonomic, genetic, and ecologic studies of present animal life form the necessary basis for interpretation of fossils (Rand, 1948, 1954). To give examples of some recent contributions, the archaeological studies by R. S. MacNeish in the Yukon (Firth River area) have supplied much useful information on fossil faunas and paleoclimate. Other rich fossil localities are known from the Yukon, Alaska, where, at Dawson, a geologic and paleontologic study is in progress by O. L. Hughes of the Geological Survey of Canada. Many contributions have been made by scientists of the Royal Ontario Museum and the National Museum of Canada, for example.

Lists on Quaternary invertebrate faunas exist from many areas. Such lists of species are certainly useful; but from the climatological as well as the paleoecologic point of view, data on the relative abundance of different species and their distribution in time, and also information on representative size and weight of individuals is much desired and would add considerably to the usefulness of such studies. This approach on molluscan faunas of fresh-water origin, presently pursued by La Roque (1960), Reynolds (1959), and Elson (1960), has clearly shown the advantages of a wider scope used in studies of marine molluscs of the Champlain Sea episode in the St. Lawrence lowlands. Elson's studies (1960) have shown that the Champlain Sea episode, lasting from 11,500 to 9500 years B.P. (Terasmae, 1960), can be divided into four phases:

(1) *Pecten* phase. Probable deep water phase.

(2) *Hiatella* phase. Subarctic sea with summer temperatures from 0° C. to 5° C. and salinity from 28 to 23‰. Probable age from 11,300 to 10,600 years B.P.

(3) *Mya arenaria* phase. Summer temperatures increasing from 5° C. to perhaps 12° C. (locally 15° to 20° C.). Salinity probably decreased from 25‰ or 20‰ as the water freshened. Probable age 10,600 to 10,000 years B.P.

(4) The *Lampsilis* lake. Brackish to fresh water environment. End of the Champlain Sea episode.

Botanical studies. Under this heading two major points deserve mention. First, many species of plants have not had their present distribution ranges in the past; and, second, the migration and existence of plants within their specific ranges are largely governed by ecologic factors that in turn are intimately related to climate. For example, a change in climate affects the ecologic site

and this either favors or limits the existence of certain plant species at that site. Depending on such changes, the plant communities change in composition. Migration of plants may occur in a direction of more favorable environments.

It is here that studies of taxonomy, ecology, genetics, history, and geography of modern plants yield pertinent information on paleoclimatology (Baldwin, 1958; Danserau, 1953; Deevey, 1949; Dillon, 1956; Dore and Gillett, 1955; Grayson, 1956; Griggs, 1937; Halliday, 1937; Hare, 1959; Horton and Bedell, 1960; Hustich, 1949; Löve and Löve, 1957; Löve, D., 1959; Löve, A., 1959; Native Trees of Canada, 1956; Polunin, 1940; Porsild, 1955, 1957, 1958; Raup, 1941; Savile, 1956; Scoggan, 1957). Extensive botanical studies are in progress by J. A. Calder in British Columbia and the Yukon by D. B. O. Savile in the Arctic Archipelago, and by J. C. Ritchie in Manitoba, to mention a few.

Discontinuous ranges of distribution of some species may indicate that such ranges were continuous in the past, and thus that climatic conditions may have prevailed that favored that particular species. An advancing or retreating tree line, for example, may indicate that climatic conditions are either improving or deteriorating (Griggs, 1937).

Certain soil profiles are associated with specific types of plant communities and, where such association does not exist, a climatic change may have caused a change in vegetation, whereas the soil profile, being more resistant to change, remains unaltered for some time. A situation where forest is invading prairie or vice versa may illustrate this point.

Study of soils. The different physiographic regions and phytogeographic regions have their distinctive climatic conditions and are characterized by differences in the soil profile that has developed. Particularly in areas that were covered or strongly influenced by Pleistocene glaciations, and where migrations of plants have taken place, changes in the soil profile also have occurred. Such changes are easier to detect in some areas than in others. In border line areas between distinct physiographic and phytogeographic regions these changes can be observed quite readily. An interesting study has been made by Sharp (1942) in the Yukon where he noticed patterned ground on the slopes in the St. Elias Range. These features occur largely above timber line and below the permanent frost line. Near and below the timber line these features are stable, as indicated by invasion by vegetation and some modification by weathering and erosion. Similar features near the snow line are active. In the intermediate area some stone stripes rest on earthy subsoil containing abundant plant roots. Sharp believes that this relation indicates a rejuvenation of soil structure development caused by the accession of cooler climate following a mild period during which frost action has been less severe.

Another interesting example has been observed by Hills (1959, and personal communication) from the Clay Belt of northern Ontario where soil drainage conditions, following the drainage of Glacial Lake Barlow-Ojibway about 6000 to 7000 years ago, were better then than later, partly owing to greater slope towards James Bay in early post-Glacial time. It is possible that during the drier and warmer hypsithermal interval, about 7500 to 5000 years ago, even fine-grained silty clay soils could have been reasonably well drained and a corresponding soil profile developed. Later, a cooling of climate or an increase in precipitation, or both, caused the development of the glei layer that is now

characteristic of the poorly drained soils of the Clay Belt. This glei layer seems to have been developed in, and superimposed on, an earlier, different soil profile.

Although such plant-soil dissociations and superimposed soil profiles or soil profiles out of harmony with their physiographic regions can be good indicators of climatic changes, it is often difficult to determine the age of such changes from the soil profiles, except in a relative sense.

Palynology and paleobotany. Palynological and paleobotanical studies help to reconstruct the past development, migration, and composition of vegetation. Paleobotanical studies of Canadian Quaternary deposits extend back about 100 years and several reports include results from these investigations (Baker, 1920; Coleman, 1941; Hinde, 1878). The main drawback in paleobotanical studies, from the paleoclimatologic point of view, is the lack of quantitative data. In most cases, only a list of plant species is given.

Palynological studies, however, have overcome this difficulty. Furthermore, plant microfossils can be found in a much greater variety of sediments, and sufficient numbers of pollen and spores, to allow statistical treatment of the results, can be recovered from small samples. These are important advantages that have been used with success in stratigraphic and paleoecologic work.

Palynological studies made of Quaternary deposits in Canada are rather widely scattered. Somewhat better coverage is available for Quebec and Ontario (some results still unpublished), where numerous studies have been made in recent years (Auer, 1930; Grayson, 1956; Ignatius, 1956; Potzger, 1953*b*; Potzger and Courtemanche, 1956; Terasmae, 1958, Wenner; 1947).

In British Columbia palynological studies have been made by Hansen (1949, 1950) and Heusser (1955), and by myself in scattered localities in that province (Terasmae and Fyles, 1959).

Palynological studies have been planned by J. C. Ritchie in Manitoba and Mrs. M. Steeves in Saskatchewan where the writer has made isolated studies of late-glacial deposits.

Palynological investigations (unpublished) made by the writer in New Brunswick, Nova Scotia, Prince Edward Island and Newfoundland have yielded much useful information, pertinent to paleoclimatologic speculation. A detailed study has been made in Cape Breton island by Livingstone and Livingstone (1958), and Ogden (1960) has studied pollen stratigraphy in Kings County, Nova Scotia.

I have made palynological studies of Quaternary beds in several areas in the Northwest Territories (Terasmae and Craig, 1958; Terasmae, 1959) but this work is still in progress.

Palynological studies that I have made in the Arctic islands are still in the preliminary stage and hence unpublished. However, these studies have shown that continuous sequences of sediments extend beyond the range of the radiocarbon dating in some areas indicating that such areas were not glaciated at least during the last glaciation. Livingstone's studies in Alaska (1955) have been very helpful for reference and comparison.

Climatic Changes

Interpretation of climatic changes (Deevey, 1949; Ewing and Donn, 1956; Flint, 1947; Landsberg, 1949) beyond the range of written records is a fasci-

nating field for speculation. The factors to be considered are numerous and the pitfalls are many, some of which are known: others not. The reliability and accuracy of paleoclimatologic studies largely depends on the alertness, experience, and ability of observation and interpretation of the scientist. Certain kinds of evidence are very obvious: for example, the existence of dry lake basins in arid regions constitute evidence of former climates with more moisture. Changes in abundance of certain species within a plant association are much more difficult to interpret from a climatologic point of view and require intimate knowledge of ecologic tolerance and requirements of the species concerned. An excellent example of such a study is the report on white and red pine by Horton and Bedell (1960). Experience has shown that interpretation of past climates should be based on as many criteria as possible and this calls for team work where scientists from several fields join their forces for mutual benefit.

The plant-animal-climate relationship is very complex, and great caution should be exercised in reaching paleoclimatologic conclusions. In such conclusions adequate attention must be given also to edaphic factors and the history of the species with its genetic potentialities and deficiencies. The geological history of the region considered is of equal importance.

Southern Ontario and Quebec. In this region evidence of late-Quaternary climates extends back about 100,000 years beginning with the interglacial Don beds at Toronto (Coleman, 1932; Hinde, 1878; Terasmae, 1960; Watt, 1953). Although an unknown length of time is not represented by deposition in the beginning of that interglacial interval (probably Sangamon), evidence found in deposits of the remaining part of that interval indicates that climatic conditions were closely similar to the present at Toronto, and toward the end of the interglacial interval a deterioration of climate occurred. It is inferred from palynological studies that the annual mean temperature was lowered by about 15° F. in the closing phase of that interglacial interval, bringing about a change from mixed hardwood forest to a boreal forest. A further lowering of the annual mean temperature by another 15° to 20° F. would have introduced tundra conditions, but satisfactory evidence for this is lacking. At present it seems more likely that the forest in the Toronto area may have been eventually overridden by the main Wisconsin ice. However, the Wisconsin ice did not advance to the limit of its maximum extent in one grand event. Halts and retreats of the ice margin occurred, as indicated by nonglacial deposits (Dreimanis, 1960*b*; Terasmae, 1960). It is probable that an increase of some 10°–15° F. in the annual mean temperature occurred in southern Ontario during these advance interstadial intervals of the Wisconsin glaciation.

In all the previous reasoning it was assumed that the annual precipitation did not decrease; rather it was the ratio of rain and snow that decreased. At present the annual precipitation at Toronto consists of 25.5 inches of rain and 54.6 inches of snow. In the boreal forest region (at Kapuskasing) this ratio is about 18.4 inches rain and 95.8 inches snow and in the central areas of Labrador (at Knob Lake) the ratio is 14.7 inches of rain and 128.6 inches of snow. In all these areas there is not too much difference in the total precipitation, measured as water. If this trend is followed and accompanied by a slight decrease in the annual mean temperature, accumulation of glacier ice will inevitably result, first in areas such as central Labrador-Ungava.

Some areas in southwestern Ontario became ice-free some time before 12,000 years ago and, possibly, 13,000 to 14,000 years B.P. as shown by radiocarbon dates. With all probability tundra conditions existed at least locally, and palynological studies I have made lend some support to this statement. From the time of deglaciation until some 10,000 years B.P. boreal forest covered southwestern Ontario. It seems reasonable to assume that climatic conditions were somewhat similar to these in the present boreal forest region of Ontario. January mean temperatures may have ranged from 0° F. to 10° F. Extreme low temperatures in the winter may have been frequently −50° F. to −60° F. Late frosts may have occurred in June, and summer temperatures reached 90° F. to 100° F. July mean temperature probably ranged in the 60s. Summers were short and frosts may have occurred in September and October. Precipitation during summer was higher than in the winter. Average total precipitation ranged from 20 to 30 inches, and winter snowfall averaged from 60 to 100 inches.

These climatic conditions quite effectively eliminated most hardwood species from the forest composition.

TABLE 1

	Pine area	Extremes in pine range
Annual average of daily mean temperature	40° F.	32 to 44° F.
January average of daily minimum temperature	−3° F.	−15 to 18° F.
July average of daily maximum temperature	79° F.	64 to 81° F.
Average total precipitation	30.2″	20.3 to 53.1″

The episode from 10,000 to 8000 years ago is characterized by a decrease in spruce and a corresponding increase in jack pine. This change could have been caused by a slight decrease in summer precipitation or an increase in summer temperature resulting in a longer, drier summer, whereas extreme cold spells in the winter still may have been frequent. It is possible that jack pine was favored under such conditions.

About 8000 to 7000 years ago the jack pine was replaced first by red pine and then by both white pine and red pine. Horton and Bedell (1960, p. 17) have described the climate within the range of white and red pine (TABLE 1). For the area with greatest pine concentration they state on page 17, ". . . the data indicate a moderate temperature, but somewhat low precipitation. Summers are warm but somewhat droughty. The abundance of the pine in the various parts of the range is controlled to a large extent by climate as it favors or discourages competing species and the incidence of fires." Furthermore, on p. 53: "Generalizing, it is apparent that the pines prefer dry conditions. Stands are abundant and relatively stable only in the climatically drier sections, and in moister sections the pine species are largely restricted to dry sites by strong competition from tolerant hardwoods or, to the north, boreal mixedwoods." These statements seem to be well substantiated by palynological studies.

Hardwood species were present in southern Ontario all through the pine period lasting from 7000 to 5000 years ago. Then a marked decrease in pine

occurred, with a corresponding increase of hemlock and beech. The study by Horton and Bedell supports the interpretation, based on palynological studies, that a change in climate at the close of the pine period involved more moisture during summer. The average annual mean temperature probably remained unchanged and about 5° F. higher than the present. Mixed hardwood forest was favored under these conditions.

About 2500 to 3000 years ago a cooling of climate took place. This change again favored the boreal species.

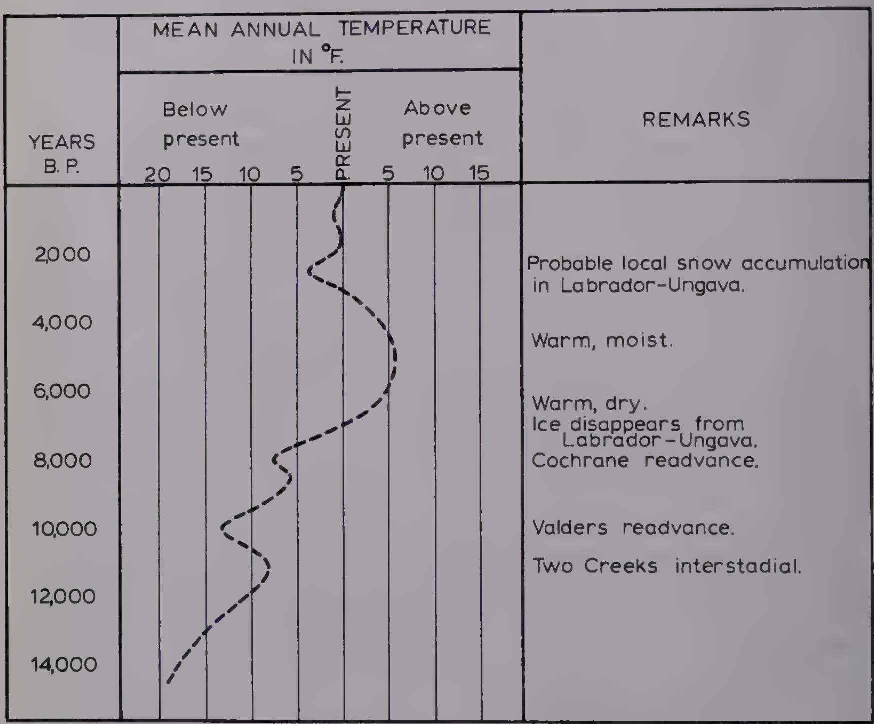


FIGURE 1. Tentative paleotemperature curve for southeastern Canada, based on palynological studies and other available data.

During the hemlock-beech period fluctuations in climate are indicated by a definitely bimodial maximum in the hemlock graph in southeastern Canada. More detailed palynological studies, coupled with radiocarbon dating, are planned to investigate these more recent climatic oscillations.

Based on the palynological evidence, coupled with other available data, an attempt has been made to construct a curve showing changes in the annual mean temperature in southeastern Canada in the last 14,000 years. This generalized curve is admittedly tentative and may have to be modified as more information accumulates (FIGURE 1).

The maritime provinces. A development of climate essentially similar to that of southern Ontario and Quebec in late- and post-Glacial time has been estab-

lished for the Maritime Provinces, based on about 10 pollen diagrams worked out by myself in that region and additional studies by Auer (1930), Livingstone and Livingstone (1958), and Ogden (1960).

Certain differences are characteristic of pollen diagrams from Newfoundland. First, one is dealing with fewer indicator species of tree pollen, commonly used by palynologists elsewhere and, second, climatic changes have been moderated by the influence of the Atlantic Ocean. However, the palynological studies made so far in Newfoundland are too few to permit a detailed discussion on paleoclimates.

Northern Ontario and Quebec. Palynological studies made in these areas are not numerous. The time covered extends back about 7000 years from the present (Grayson, 1956; Ignatius, 1956; Potzger and Courtemanche, 1956; Terasmae and Hughes, 1960; Wenner, 1947). I have found evidence in the basal parts of pollen diagrams for a climate probably warmer than the present. This opinion is in agreement with other studies from the region, and it seems logical to expect that the remnants of the continental ice melted during the hypsithermal interval. The annual mean temperature then was about 5° F. higher than the present; summer and, possibly, also winter precipitation was probably lower. The hypsithermal interval came to a close when the average annual mean temperature gradually dropped some 10° F. and was accompanied by heavier accumulation of snow. The climate then was more severe than it is now and may have resulted in expansion of the permafrost area. Supporting evidence has been found in the James Bay lowland (Baldwin and Porsild, personal communication).

Present climatic conditions in Labrador-Ungava and northern Quebec deserve special mention because this region probably was the area of initial accumulation of snow gradually leading to formation of the continental ice sheet.

The vast area east of Hudson and James Bays extending to Hudson Strait in the north, to the coast of Labrador in the east, and to the Gulf of St. Lawrence in the south is largely a highland plateau with a monotonous surface strewn with lakes and muskeg. Winters in this region are long and cold. At Schefferville (Knob Lake) the annual mean temperature is about 24° F., with January mean ca. −12° F. Summers are cool, but warm days may occur, with temperatures reaching 90° F. The frost-free season varies from 40 to 80 days. A large section of this region has permanently frozen subsoil, chiefly north of latitude 55° N. Precipitation varies considerably with topography and elevation, averaging from about 40 inches in the interior to 15 inches to the west and north. Snowfall is heavy, more than 100 inches in the interior, and total accumulation of snow on the ground at the end of winter exceeds four feet.

It is quite possible that a slight lowering of average summer temperatures, with precipitation remaining unchanged, will allow accumulation of perennial snowbanks at suitable sites. If such lower mean temperatures, about 5° F. to 10° F. below the present, persist over a number of years continuous snowfields may form over large areas in Labrador-Ungava. Thus it seems plausible that even relatively small changes in temperature and, perhaps, in precipitation can

initiate climatic conditions which, if prevailing over a period of time, may well become the beginning of another glaciation.

Western plains region. Palynological studies in this region are entirely too few to allow reconstruction of a climatic sequence for the postglacial time. However, the available evidence indicates that the retreat of ice was followed, perhaps after a short tundra episode, by the boreal forest and not directly by a prairie vegetation. In late-Glacial time, 12,000 to 10,000 years ago, some areas in southern Saskatchewan were covered by boreal forest (Kupsch, 1960; my own unpublished studies).

Palynological studies in the extensive prairie region will encounter certain specific difficulties because of the largely treeless vegetation and interpretation of the pollen and spore record.

West Coast and interior British Columbia. Studies made thus far have indicated that a paleoclimatologic sequence can be worked out for the West Coast (Hansen, 1950; Heusser, 1955; Terasmae and Fyles, 1959). Precipitation seems to have been adequate or excessive at all times during the post-Glacial period, and temperature is the controlling factor although even this is moderated by the influence of the Pacific Ocean. Hansen has demonstrated climatic changes in post-Glacial time on the West Coast, and it appears that a similar climatic sequence to that of eastern North America also applies here. A post-Glacial warmer and drier period on the West Coast seems to be the correlative of the hypsithermal interval.

Conditions in the interior British Columbia are entirely different, and precipitation is the limiting factor in the valleys and on the plateaux. Further difficulties are introduced by the markedly different vegetation on the north- and south-facing slopes of the mountains. It is not clear how a climatic change will influence the sum total of pollen and of spores derived from the north and south slopes and from the valley bottom in any particular locality.

Arctic Canada. Preliminary studies have shown that much more information pertaining to late-Quaternary climates can be found in the Arctic than had been expected previously. Until a few years ago the Canadian Arctic was still largely unexplored. The recent economical and political interest in the far north has opened up an entirely new episode for the Arctic. This increased activity has enabled the scientist to gain access to remote islands in the Arctic Archipelago and to northern coastal areas of the mainland. Large scale geologic reconnaissance by the Geological Survey of Canada (1959) has provided the necessary basic coverage, and a number of separate research teams have carried on field work in the Arctic. Most of these studies, however, are still in progress and only preliminary reports have been issued.

Meteorologically the Canadian Arctic is still almost unexplored. Weather records from the stations in the far north are available for the last few years only; hence they may or may not be representative of a longer period of time. This is unfortunate, for interpretation of paleoclimates is very much dependent on the present plant-animal-soil-climate relationships. This lack of reference data on present climate is felt more in the Arctic than elsewhere because marked differences between physiographic regions and from season to season have been noticed. Furthermore, plants and animals in the Arctic Archipelago are

existing in environments where a small change in climate means the difference between survival and extinction.

The Arctic is a land of contrasts, both real and imaginary. The view that the Arctic is a land of snow and ice is true in a sense, yet it is a true rock desert at the same time. Annual average precipitation is extremely low; only 2 to 3 inches in northern Arctic Archipelago and less than 10 inches in southern parts of the Archipelago. The Arctic climatic region is that part of Canada north of the July 50° F. isotherm, which runs from MacKenzie delta to Churchill, Manitoba, around the southern Hudson Bay and down the Labrador coast. This isotherm corresponds in general to the northern limit of tree growth. The extremely long nights in winter and days in summer are an important factor in climatic control, and this latitudinal control, involving also incoming radiation coupled with areal distribution of land and water, determines in a general way the climatic limits within the Arctic region. In addition to the MacKenzie Basin, the Arctic climatic region includes the Barrens and the Arctic Archipelago. The Barrens comprise the part of Northwest Territories that stretches northward to the Arctic Ocean from the northern limit of tree growth and lies between the Hudson Bay and the MacKenzie delta area. Countless lakes, swamps, muskeg, and rock outcrop characterize this low-lying region. Summers are very short and cool. The average frost-free period is 40 to more than 60 days although freezing temperatures can occur during any month of the year. July and August are the only months with an average temperature above 40° F. Winters are extremely long and cold. Precipitation averages from 7 to 8 inches in the north to about 12 inches in the southern part of the Barrens. Snowfall contributes about one half of the total precipitation. Clear skies characterize the winter months along the Arctic coast, but late summer and early fall are cloudy, accompanied by the precipitation maximum at that time. Generally strong winds comprise another important climatic factor in the Barrens.

It is interesting to note that although the mean annual temperatures are lower in the Arctic Archipelago than anywhere else in Canada, ranging from 15° F. to 0° F. on Melville and Ellesmere Islands, the extremes are not as severe as they would be in continental areas of the same latitude. Summer temperatures are kept uniformly cool by the ice-filled polar waters, with surface temperatures near 30° F., which prevent the air in contact with them from warming. On the other hand, winter temperatures are probably slightly moderated by radiation from the water beneath the ice cover, which is usually less than 6 feet in thickness.

The occurrence of occasional mild periods during the Arctic winter deserves mention. The intense cyclonic activity in the Davis Strait area can bring the relatively warm air from the Atlantic over the eastern Arctic. According to weather records one such warm spell occurred in January 1958, when the mean temperatures for the north were some 15° F. above normal in northern Ellesmere Island. These mild spells usually do not extend west of Cornwallis and Somerset Islands.

It is possible, as a suggestion, that such abnormal circulation conditions may contribute somewhat to the moisture regime on the large eastern islands of the

Arctic Archipelago in order to maintain small ice caps on Baffin, Devon, and Ellesmere Islands, which lie in a high mountain range with peaks rising above 10,000 feet.

A very interesting and important problem in Quaternary research has been the existence of unglaciated areas in the far north. Views both for and against the existence of such areas have been presented, and it is interesting to note that the term "unglaciated" may have a different meaning depending on whether it is used in a geologic or biological sense. In the first case, unglaciated areas are those that were not covered by glaciers and where there is no evidence of erosion or deposition by ice. Biologically the term generally means that plant and animal life could have existed in such areas. Thus it is possible that an area shown as unglaciated on a geologic map was glaciated in a biological sense. A low summer temperature and a frequently overcast sky with generally high winds, coupled with cold winters and low precipitation can produce conditions where a growth period for plants is almost nonexistent, and snow covers the more sheltered sites, whereas exposed sites may not support any plant life.

Certain areas may have escaped all or at least some of the glaciations, and records, coupled with those from interglacial deposits, can be discovered there of past plant and animal life. Interglacial pollen and spore-bearing deposits have been discovered already on several of the Arctic islands. Studies of these deposits are in progress and as yet unpublished. Observations in the field (J. G. Fyles, personal communication) and palynological studies have shown that in interglacial intervals trees were growing on Banks Island, and appreciable thickness of peat has accumulated in several areas on the island in late-Quaternary time.

On the mainland, palynological studies have shown that the distribution ranges of some plant species extended farther north about 5000 to 6000 years ago, indicating climatic conditions more favorable than the present (Terasmae and Craig, 1958). Alluvial and lacustrine sediments in the Thelon Valley, Northwest Territories, in which plant fossils were found, dated at 5400 $\pm$ 230 years B.P. (L-428), have been frost-heaved, and they form a pingo indicating a deterioration of climate later than about 5000 years ago.

When similar geologic observations become more numerous and many more palynological and paleobotanical studies, supported by radiocarbon dating are made, it will perhaps be possible to compile a climatic sequence covering most of the Quaternary epoch for the Canadian Arctic. It will be possible to achieve this only when information from all sources is compiled and after such data have been critically examined and interpreted by scientists in both the physical and biological fields.

Time Equivalence of Late-Quaternary Climatic Changes

Time equivalence of climatic changes in the Quaternary epoch is the fundamental requirement on which all correlations using palynological and paleontologic evidence are based. Recent studies seem to indicate that this is true probably on a world-wide basis, although more studies are required to substantiate the statement. At least within the limits of any one continent the climatic changes appear to satisfy the requirement of time equivalence.

However, terms such as late-Glacial, post-Glacial and Recent are necessarily relative depending on the geographical location. It is obvious that the late-Glacial episode becomes younger in a south to north direction following the retreat of the ice margin. This presents some difficulties in correlating, for example, the late-Glacial deposits of Wisconsin age. When the plant communities followed and colonized the area uncovered by the retreating ice margin it is possible that similar fossil assemblages, but of different age, were to be found in two localities separated by some distance in a north-south direction. Radiocarbon age determinations have helped to overcome these problems.

The problem of migration rates for plants and animals is of major importance in paleoclimatological interpretations (Savile, 1956). For example, in Canada many species must have covered hundreds and even thousands of miles in late- and post-Glacial time in order to occupy their present distribution ranges. Studies made of fossils in late-Quaternary beds seem to indicate that these migrations occurred rapidly and, perhaps, in a haphazard manner for several species which, in some measure, may account for the present discontinuous ranges. However, the demonstrated ability of many species to migrate over long distances rapidly may make less convincing the argument for the existence of refugia, that is, isolated areas in which plants could survive the glaciation. Such refugia certainly did exist in Alaska and in Peary Land (Northern Greenland) in the far north. Further studies of plant and animal fossils in such areas will certainly contribute to our knowledge of paleoclimates of late-Quaternary time.

References

Antevs, E. 1925. Retreat of the last ice-sheet in Eastern Canada. Geol. Survey Can. Mem. **146**.

Antevs, E. 1931. Late-glacial correlations and ice recession in Manitoba. Geol. Survey Can. Mem. **168**.

Armstrong, J. E. 1957. Surficial geology of New Westminster map-area, British Columbia. Geol. Survey Can. Paper 57-5.

Auer, V. 1930. Peat bogs in southeastern Canada. Geol. Survey Can. Mem. **162**.

Baker, F. C. 1920. The life of the Pleistocene or glacial period. Ill. Univ. Bull. **17**(41).

Baldwin, W. K. W. 1958. Plants of the Clay Belt of northern Ontario and Quebec. Natl. Museum Can. Bull. **156**.

Bird, J. B. 1959. Recent contributions to the physiography of northern Canada. Z. Geomorph. **3**(2): 151–174.

Canada Year Book. 1960. The climate of Canada. Met. Branch, Air Service, Dept. Transp. Can.

Chapman, J. D. 1952. The climate of British Columbia. Proc. Fifth Brit. Columbia Natl. Research Conf.

Chapman, L. J. 1938. The climate of southern Ontario. Can. Geogr. J. **17**(3): 136–141.

Chapman, L. J. & D. F. Putnam. 1951. The Physiography of Southern Ontario. Univ. Toronto Press. Toronto, Ont., Canada.

Chapman, L. J. 1953. The climate of northern Ontario. Can. J. Agric. Sci. **33**(1): 41–73.

Coleman, A. P. 1933. The Pleistocene of the Toronto Region. Ont. Dept. Mines Rept. **41**(7): 1–55.

Coleman, A. P. 1941. The Last Million Years. Univ. Toronto Press. Toronto, Ont., Canada.

Cook, F. A. 1960. Selected bibliography on periglacial phenomena in Canada. Can. Dept. Min. Techn. Surveys, Geogr. Br. Bibliogr. Ser. 24.

Craig, B. G. 1956. Surficial geology of the Drumheller area, Alberta, Canada. Unpubl. Ph.D. Thesis. Univ. Michigan. Ann Arbor, Mich.

Craig, B. G. 1960. Surficial geology of North-Central District of MacKenzie, Northwest Territories. Geol. Survey Can. Paper 60-18.

CRAIG, B. C. & J. G. FYLES. 1960. Pleistocene geology of Arctic Canada. Geol. Survey Can. Paper 60-10.
DANSERAU, P. 1953. The postglacial pine period. Roy. Soc. Can. Trans. 47. Ser. **3**(5): 23–38.
DEANE, R. E. 1950. Pleistocene geology of the Lake Simcoe district, Ontario. Geol. Survey Can. Mem. **256**.
DEEVEY, E. S. 1949. Biogeography of the Pleistocene. Bull. Geol. Soc. Am. **60**(9): 1315–1416.
DILLON, L. S. 1956. Wisconsin climate and life zones in North America. Science. **123** (3188): 167–176.
DORE, W. G. & J. M. GILLETT. 1955. Botanical survey of the St. Lawrence Seaway area in Ontario. Can. Dept. Agric. Sci. Serv. Rept. (mimeographed).
DREIMANIS, A. 1960*a*. Finite radiocarbon dates of the Port Talbot interstadial deposits in southern Ontario. Science. **131**(3415): 1738–1739.
DREIMANIS, A. 1960*b*. Pre-classical Wisconsin in the eastern portion of the Great Lakes region, North America. Proc. Intern. Geol. Congr. Norden. Pt. IV. Sect. **4**: 109–119.
ELSON, J. A. 1955. Surficial geology of the Tiger Hills region, Manitoba. Unpub. Ph.D. Thesis. Yale Univ. New Haven, Conn.
ELSON, J. A. 1957. Lake Agassiz and the Mankato-Valders problem. Science. **126**(3281): 999–1002.
ELSON, J. A. 1959. Classification and chronology of the Pleistocene. *In* Problems of the Pleistocene and Arctic. Publ. McGill Univ. Museums. **1**: 1–25.
ELSON, J. A. 1960. Littoral mollusks of the Champlain Sea. Excursion guide. Dept. Geol. Sci. McGill Univ. Montreal, Canada.
EWING, M. & W. L. DONN. 1956. A theory of ice ages. Science. **123**(3207): 1061–1066.
FAIRBRIDGE, R. W. 1958. Dating the latest movements of the Quaternary sea level. Trans. N.Y. Acad. Sci. Ser. II. **20**(6): 471–482.
FAIRBRIDGE, R. W. 1960. The changing level of the sea. Scientific American. **202**(5): 70–79.
FLINT, R. F. 1947. Glacial Geology and the Pleistocene Epoch. Wiley. New York, N.Y.
FYLES, J. G. 1956. Surficial geology of the Horne Lake and Parksville map-areas, Vancouver Island, British Columbia. Dissert. Abstr. **16**(11): 2136.
GADD, N. R. 1960. Surficial geology of the Becancour map-area, Quebec. Geol. Survey Can. Paper 59–8.
GEOLOGICAL SURVEY OF CANADA. 1959. Helicopter operations of the Geological Survey of Canada. Geol. Survey Can. Bull. **54**.
GLACIAL MAP OF CANADA. 1958. Geol. Assoc. Can. Toronto, Ont., Canada.
GOLDTHWAIT, R. P. 1960. Dating the Little Ice Age in Glacier Bay, Alaska. Inst. Polar Stud. Ohio State Univ. (Mim. manuscr.).
GRAVENOR, C. P. 1957. Surficial geology of the Lindsay-Peterborough area, Ontario. Geol. Survey Can. Mem. **288**.
GRAYSON, J. F. 1956. Post-glacial history of vegetation and climate in the Labrador-Quebec region as determined by palynology. Unpub. Ph.D. Thesis. Univ. Michigan. Ann Arbor, Mich.
GRIGGS, R. F. 1937. Timberlines as indicators of climatic trends. Science. **85**(2202): 251–255.
HALLIDAY, W. E. D. 1937. A forest classification for Canada. Can. Dept. Min. Res., Forestry Br. Domin. Forest Serv. Bull. **89**: 1–50.
HANSEN, H. P. 1949. Postglacial forests in south central Alberta, Can. Am. J. Botany. **36**(1): 54–65.
HANSEN, H. P. 1950. Pollen analysis of three bogs on Vancouver Island, Canada. J. Ecol. **38**(2).
HARE, F. K. 1950. The climate of the eastern Canadian Arctic and Subarctic and its influence on accessibility. Ph.D. Thesis. Univ. Montreal. Montreal, Que., Canada.
HARE, F. K. 1952*a*. The climate of the Island of Newfoundland: a geographical analysis. Geogr. Bull. **2**: 36–88.
HARE, F. K. 1952*b*. Post-glacial climatic change in Eastern Canada. Can. Br. Roy. Met. Soc. Papers. **2**(7).
HARE, F. K. 1959. A photo-reconnaissance survey of Labrador-Ungava. Can. Dept. Min. Techn. Surveys, Geogr. Branch, Mem. 6.
HARLOW, W. H. & E. S. HARRAR. 1941. Textbook on Dendrology. McGraw-Hill. New York, N. Y.
HATTERSLEY-SMITH, G. 1958. Glaciological research in northern Ellesmere Island. Can. Geographer. **12**: 32–34.

HATTERSLEY-SMITH, G. 1959. Research in the Lake Hazen region of northern Ellesmere Island in the International Geophysical Year. Arctic Circular. **12**(1): 2–12.

HENDERSON, E. P. 1959*a*. Surficial geology of Sturgeon Lake map-area, Alberta. Geol. Survey Can. Mem. **303**.

HENDERSON, E. P. 1959*b*. A glacial study of central Quebec-Labrador. Geol. Survey Can. Bull. **50**.

HENDERSON, E. P. 1959*c*. Surficial geology of St. John's, Newfoundland. Geol. Survey Can. Map 35-1959.

HEUSSER, C. J. 1955. Pollen profiles from the Queen Charlotte Islands, British Columbia. Dept. Expl. & Field Research. Am. Geogr. Soc. New York (manuscript).

HIBBARD, C. W. 1949. Pleistocene vertebrate paleontology in North America. Bull. Geol. Soc. Am. **60**(9): 1417–1428.

HILLS, G. A. 1959. A ready reference to the description of the land of Ontario and its productivity. Ont. Dept. Lands Forests, Div. Res. Prelim. Rept.

HINDE, G. J. 1878. Glacial and interglacial strata of Scarborough Heights. J. Can. Inst. (Roy. Can. Inst.).

HORTON, K. W. & G. H. D. BEDELL. 1960. White and red pine—ecology, silviculture and management. Can. Dept. North. Affairs Natl. Resources. Forestry Branch Bull. **124**.

HUGHES, O. L. 1956. Surficial geology of Smooth Rock Cochrane District, Ontario. Geol. Survey Can. Paper 55-41.

HUSTICH, I. 1949. On the forest geography of the Labrador peninsula. A preliminary synthesis. Acta Geogr. **10**.

IGNATIUS, H. G. 1956. Late-Wisconsin stratigraphy in North-Central Quebec and Ontario, Canada. Ph.D. Thesis. Yale Univ. New Haven, Conn.

IVES, J. D. 1959*a*. The deglaciation of the Helluva Lake area, 50–70 miles northwest of Schefferville. McGill Subarctic Research Papers. **6**: 45–49.

IVES, J. D. 1959*b*. Glacial drainage channels as indicators of late-glacial conditions in Labrador-Ungava: a discussion. Cahiers de Geogr. de Quebec. **3**(5): 57–72.

IVES, J. D. 1960. The deglaciation of Labrador-Ungava—an outline. Cahiers de Geogr. de Quebec. **4**(8): 323–343.

JOHNSTON, W. A. 1917. Pleistocene and recent deposits in the vicinity of Ottawa, with a description of the soils. Geol. Survey Can. Mem. 101.

KARLSTROM, T. N. V. 1956. The problem of the Cochrane in late Pleistocene chronology. U. S. Geol. Survey Bull. 1021-J.

KARLSTROM, T. N. V. 1960. The Cook Inlet, Alaska, glacial record and Quaternary classification. U.S. Geol. Survey Prof. Paper. **400-B**: 330–332.

KARROW, P. F. 1957. Pleistocene geology of the Grondines map-area, Quebec. Ph.D. Thesis. Univ. Ill. Urbana, Ill.

KARROW, P. F. 1959. Pleistocene geology of the Hamilton map-area. Ont. Dept. Min. Geol. Circular 8.

KENDREW, W. G. & B. W. CURRIE. 1955. The Climate of Central Canada. Queen's Printer. Ottawa, Ont., Canada.

KENDREW, W. G. & D. KERR. 1955. The Climate of British Columbia and the Yukon Territory. Queen's Printer. Ottawa, Ont., Canada.

KOEPPE, C. E. 1931. The Canadian Climate. McKnight & McKnight. Bloomington, Ill.

KUPSCH, W. O. 1960. Radiocarbon-dated organic sediment near Herbert, Saskatchewan. Am. J. Sci. **258**(4): 282–292.

LANDSBERG, H. 1949. Climatology of the Pleistocene. Bull. Geol. Soc. Am. **60**(9): 1437-1442.

LA ROQUE, A. 1960. Quantitative methods in the study of non-marine Pleistocene mollusca. Proc. Intern. Geol. Congr. Norden. Pt. IV. Sect. **4**: 134–141.

LAVERIERE, C. & A. COURTEMANCHE. 1959. La geomorphologie glaciaire de la region du Mont Tremblant. Revue Can. Geogr. **13**(3–4): 103–134.

LEE, H. A. 1955. Surficial geology of Edmundston, Madawaska and Temiscouata Counties, New Brunswick and Quebec. Geol. Survey Can. Paper 55-15.

LEE, H. A. 1957. Surficial geology of Fredericton, York and Sunbury Counties, New Brunswick. Geol. Survey Can. Paper 56-2.

LEE, H. A. 1959. Surficial geology of southern District of Keewatin and the Keewatin ice divide, Northwest Territories. Geol. Survey Can. Bull. 51.

LEE, H. A. 1960. Late glacial and postglacial Hudson Bay Sea episode. Science. **131** (3413): 1609–1611.

LIVINGSTONE, D. A. 1955. Some pollen profiles from Arctic Alaska. Ecology. **36**(4): 587–600.

LIVINGSTONE, D. A. & B. G. R. LIVINGSTONE. 1958. Late-glacial and postglacial vegeta-

tion from Gillis Lake in Richmond County, Cape Breton Island, Nova Scotia. Am. J. Sci. **256:** 341–359.

Löve, A. 1959. Origin of the Arctic flora (*in* Problems of the Pleistocene and Arctic). McGill Univ. Museums Publ. **1:** 82–95.

Löve, A. & D. Löve. 1957. Arctic polyploidy. Proc. Genet. Soc. Can. **2:** 23–27.

Löve, D. 1959. The postglacial development of the flora of Manitoba: a discussion. Can. J. Botany. **37:** 547–585.

MacClintock, P. & J. Terasmae. 1960. Glacial history of Covey Hill. J. Geol. **68**(2): 232–241.

MacKay, J. R. 1958. The Anderson River map-area, N. W. T. Can. Dept. Min. Techn. Surveys, Geogr. Branch. Mem. **5**.

MacLean's Magazine. 1960. An interim report on the next Ice Age. Dec. 3 issue. Toronto, Ont., Canada.

Mathews, W. H. 1951. Historic and pre-historic fluctuations of alpine glaciers in the Mount Garibaldi map-area, southwestern British Columbia. J. Geol. **59:** 357–380.

Meldegaard, J. 1960. Origin and evolution of Eskimo cultures in the Eastern Arctic. Can. Geogr. J. **60**(2): 64–75.

Mirynech, E. 1961. Surficial geology of the Trenton-Campbellford map-areas, Ontario. Ph.D. Thesis. Univ. Toronto. Toronto, Ont., Canada.

Native Trees of Canada. 1956. Can. Dept. North. Affrs. Natl. Resources. Forestry Branch. Bull. 61.

Nicholson, N. L. 1960. The Northwest Territories—geographical aspects. Can. Geogr. J. **60**(1): 2–27.

Ogden, J. G., III. 1960. Recurrence surfaces and pollen stratigraphy of a postglacial raised bog, Kings County, Nova Scotia. Am. J. Sci. **258:** 341–353.

Polunin, N. 1940. Botany of the Canadian Eastern Arctic, Pt. I. Natl. Mus. Can. Bull. **92**.

Porsild, A. E. 1955. The vascular plants of the western Canadian Arctic Archipelago. Natl. Museum Can. Bull. **135**.

Porsild, A. E. 1957. Illustrated flora of the Canadian Arctic Archipelago. Natl. Mus. Can. Bull. **146**.

Porsild, A. E. 1958. Geographical distribution of some elements in the flora of Canada. Geogr. Bull. **11:** 57–77.

Potzger, J. E. 1953*a*. History of forests in the Quetico-Superior country from fossil pollen studies. J. Forestry. **51:** 560–565.

Potzger, J. E. 1953*b*. Nineteen bogs from southern Quebec. Can. J. Botany. **31:** 383–401.

Potzger, J. E. & A. Courtemanche. 1956. A series of bogs across Quebec from the St. Lawrence Valley to James Bay. Can. J. Botany. **34**(4): 473–500.

Prest, V. K. 1957. Pleistocene geology and surficial deposits. Chapter 8 *in* Geology and Economic Minerals of Canada. Geol. Survey Can. Econ. Geol. Ser. 1.

Rae, R. W. 1951. Climate of the Canadian Arctic Archipelago. Can. Dept. Transp. Toronto, Ont., Canada.

Rand, A. L. 1948. Mammals of the Eastern Rockies and Western Plains of Canada. Natl. Museum Can. Bull. **108**.

Rand, A. L. 1954. The ice age and mammal speciation in North America. Arctic. **7:** 31–35.

Raup, H. M. 1941. Botanical problems in boreal America I. Bot. Rev. **7**(**3**): 147–208.

Reynolds, M. B. 1959. Pleistocene molluscan faunas of the Humboldt deposit, Ross County, Ohio. Ohio J. Sci. **59:** 152–166.

Savile, D. B. O. 1956. Known dispersal rates and migratory potentials as clues to the origin of the North American biota. Am. Midland Natl. **56**(2): 434–453.

Scoggan, H. J. 1957. Flora of Manitoba. Natl. Museum Can. Bull. **140**.

Sharp, R. P. 1942. Soil structures in the St. Elias Range, Yukon Territory. J. Geomorphol. **5:** 275–301.

Shenfeld, L. & D. F. A. Slater. 1960. The climate of Toronto. Met. Branch. Dept. Transp. Can. Circular. **3352**.

Shepard, F. P. & H. E. Suess. 1956. Rate of postglacial rise of sea level. Science. **123** (3207): 1082–1083.

Stalker, A. M. 1960. Surficial geology of the Red Deer-Stettler map-area, Alberta. Geol. Survey Can. Mem. **306**.

Terasmae, J. & B. G. Craig. 1958. Discovery of fossil *Ceratophyllum demersum* L. in Northwest Territories, Canada. Can. J. Botany. **36:** 567–569.

Terasmae, J. 1958. Contributions to Canadian palynology. Pt. 1. The use of palynological studies in Pleistocene stratigraphy; Pt. 2. Non-glacial deposits in the St. Law-

rence Lowlands, Quebec; Pt. 3. Non-glacial deposits along Missinaibi River, Ontario. Geol. Survey Can. Bull. **46**.

Terasmae, J. 1959. Palaeobotanical study of buried peat from the MacKenzie River delta area, Northwest Territories. Can. J. Botany. **37:** 715–717.

Terasmae, J. & J. G. Fyles. 1959. Palaeobotanical study of late-glacial deposits from Vancouver Island, British Columbia. Can. J. Botany. **37:** 815–817.

Terasmae, J. 1960. Contributions to Canadian palynology, 2. Part. 1. A palynological study of postglacial deposits in the St. Lawrence lowlands. Part. 2. A palynological study of the Pleistocene interglacial beds at Toronto, Ontario. Geol. Survey Can. Bull. **56**.

Terasmae, J. & O. L. Hughes. 1960*a*. Notes on glacial retreat in the North Bay area, Ontario. Science. **131**(3411)**:** 1444–1446.

Terasmae, J. & O. L. Hughes. 1960*b*. A geological and palynological study of Pleistocene deposits in the James Bay lowlands, Ontario. Geol. Survey Can. Bull. **62**.

Thomas, M. K. 1953. Climatological atlas of Canada. Natl. Research Council. Div. Bldg. Res. & Can. Dept. Transp. Met. Div., N.R.C. No. 3151.

Wagner, F. J. E. 1959. Palaeoecology of the marine Pleistocene faunas of southwestern British Columbia. Geol. Survey Can. Bull. **52**.

Watt, A. K. 1953. Glacial geology of the Toronto-Orangeville area, Ontario. Guide Book, Field Trip 3, Geol. Soc. Am. Meetings. Toronto, Ontario.

Wenner, C.-G. 1947. Pollen diagrams from Labrador. Geogr. Ann. **29:** 137–374.

THE QUATERNARY CLIMATIC CHANGES OF NORTHERN SOUTH AMERICA

Th. van der Hammen
Geological Institute, Leiden, The Netherlands

Pollen analytical studies of series of lake sediments of Pleistocene and Holocene age from the Colombian Andes have brought to light interesting facts about their climatic history.

The most important results were obtained with the upper part of a section from the Sabana de Bogotá, Colombia, the flat bottom of an actually dry Pleistocene lake, at 2560 m. above sea level and 5° north of the equator. The complete section is 200 m. thick and will probably represent the whole Pleistocene. The results of the pollen analysis of the upper 33 m. were published a short time ago (van der Hammen and Gonzalez, 1960*a*).

In addition to this long section, a number of Holocene late-glacial lake sections were analyzed from smaller lakes in different regions in the higher parts of the Colombian Eastern Cordillera. Part of the results were published (van der Hammen and Gonzalez, 1960*b*); another part will be published later.

We shall consider first the results of all these studies, and then compare them with what we know from other parts of the world. Then we shall see what is known about climatic changes in the tropical lowland of northern South America and, finally, make some general conclusions from the total available data.

Data From the Andes

From the pollen diagrams we may read the changes of vegetation. In our case these changes are of composite nature. On the one hand they represent changes between an open high-Andean vegetation type (*paramo*) and the Andean forest and, on the other hand, changes between two different types of Andean forest; the Quercetum or oak forest (and related vegetation), and the Weinmannietum (and related vegetation). The first-mentioned changes represent vertical movements of the "tree line" (or better "forest limit"); the second-mentioned represent fluctuations of the "humidity" of the climate. We may therefore easily calculate from the data of the pollen diagrams a curve indicating relative vertical fluctuations of the tree line and a curve indicating relative changes of humidity (rainfall). However, there is one complication that makes the final interpretation more difficult. Studies of the vegetation in relation to tree line and climate have shown that the altitude of the tree line depends on the humidity of the climate. Higher and more continuous rainfall and cloudiness result in a higher tree line. It will be clear that the curve indicating the relative fluctuations of this line depends on the variation of two factors: fluctuations of temperature, and fluctuations of humidity. By means of comparison with recent pollen spectra at different altitudes and in regions with different forest types, the two-mentioned relative curves could be converted into absolute ones, indicating in both cases tree-line fluctuations. With the aid of these data, a third curve may be calculated indicating fluctuations

of the tree line due to temperature only, directly convertible into a temperature curve. The three curves are drawn in FIGURE 1.

From C-14 dates we know that the uppermost 2½ m. are Holocene and that, from 2½ m. downward, the sediments are of Pleistocene age. The rate of

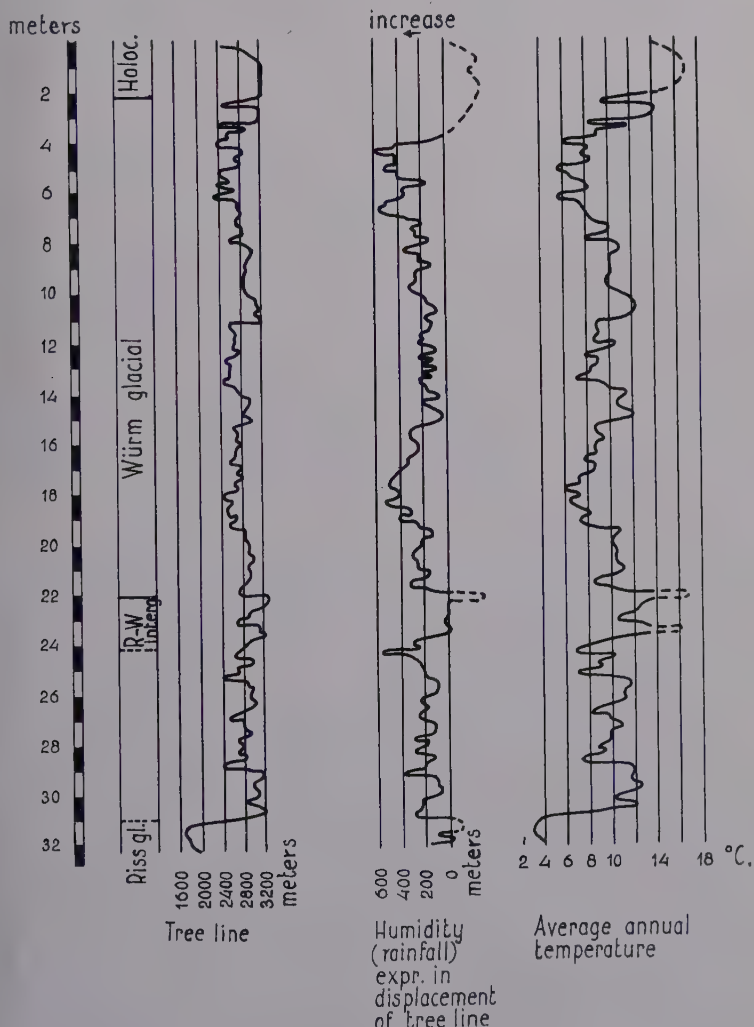


FIGURE 1. Curves of tree line, humidity, and temperature for the Upper Pleistocene in the Sabana de Bogotá, based on a pollen diagram of 33 m. of lake sediments.

sedimentation for the uppermost part was 25 to 28 cm. per 1000 years. This value agrees remarkably well with the 1 foot (±30 cm.) per 1000 years found by K. H. Clisby (unpublished communication) for the uppermost part of the sediments of the intermontane basin of the San Augustin Plains in New Mexico.

If we suppose a constant rate of sedimentation and use the values found per 1000 years to bring the curves provisionally on a time scale, we become aware of a surprising fact. The temperature curve agrees remarkably well with Emiliani's (1955) curve of ancient sea-water temperatures (Fairbridge, 1960), and with Gross's (1958) temperature curve for the Upper Pleistocene of Europe (FIGURE 2). The agreement is so complete that we cannot doubt the interpretation:

0–2,5 m.	Holocene
2,5–22,5 m.	Würm-glacial
22,5–24(29) m.	Riss-Würm interglacial (Eemian)
24(29)–33 m.	Riss-glacial

An important fact may now be deduced on comparing our temperature and humidity curves: the most important cold phases are wet phases at the same time, and the warmer phases dry. In other words the glacials are pluvials at the same time, and the interglacials are interpluvials, in this equatorial region.

The second fact we may deduce on comparing these curves with those of Emiliani and Gross is that the glacials and interglacials and also the "interstadials" and "stadials" of equatorial South America correspond in time apparently quite exactly with those established in other parts of the world (see Maarleveld and van der Hammen, 1959, for comparison with those of South Africa).

Let us now consider the minor climatic fluctuations of the Holocene.

In the Sabana de Bogotá we can clearly distinguish an alternation of climatic phases in the Holocene.

These phases are the following: VIII, wet; VII, dry; VI, wet; V, dry and warmer; and IV, dry and colder.

Several C-14 dates carried out in Copenhagen, Denmark (van der Hammen and Gonzalez, 1960*a* and *b*), indicate that zone V must correspond to the European Boreal, and that zones IV and VIII correspond probably to the pre-Boreal and sub-Atlantic.

Additional and more definite information was obtained from a number of pollen diagrams and C-14 dates (carried out in Washington) from Paramo de Palacio, in the Colombian Eastern Cordillera (van der Hammen and Gonzalez, 1960*b*). This region lies about 1000 m. higher than the Sabana de Bogotá. The climatic phases distinguished here correspond apparently in time with those of the Sabana de Bogotá. Zone VIII is relatively cold, and began around 700 to 900 B.C. Zone VII is warmer, and began shortly before 2800 B.C., while zone VI began shortly before 5300 B.C. With these data there seems no longer any doubt about the correspondence in time of the Holocene climatic fluctuations of equatorial South America with those of Europe: Boreal, Atlantic, sub-Boreal, and sub-Atlantic.

There is evidence, moreover, that the well-known late-glacial fluctuations of temperature in our region correspond equally with those in the temperate zones of the Northern and Southern hemispheres.

Interesting finally is a short and somewhat colder phase in the middle of

the Holocene, found not only in the diagrams of Paramo de Palacio, but also in a still unpublished diagram from Laguna de los Bobos, much farther north, in the department of Boyacá. This more-than-local phenomenon cannot yet be correlated with certainty with other parts of the world. We may now summarize the above-mentioned results in the following points.

(1) In the Colombian equatorial Andes, a glacial is at the same time a pluvial; an interglacial, an interpluvial.

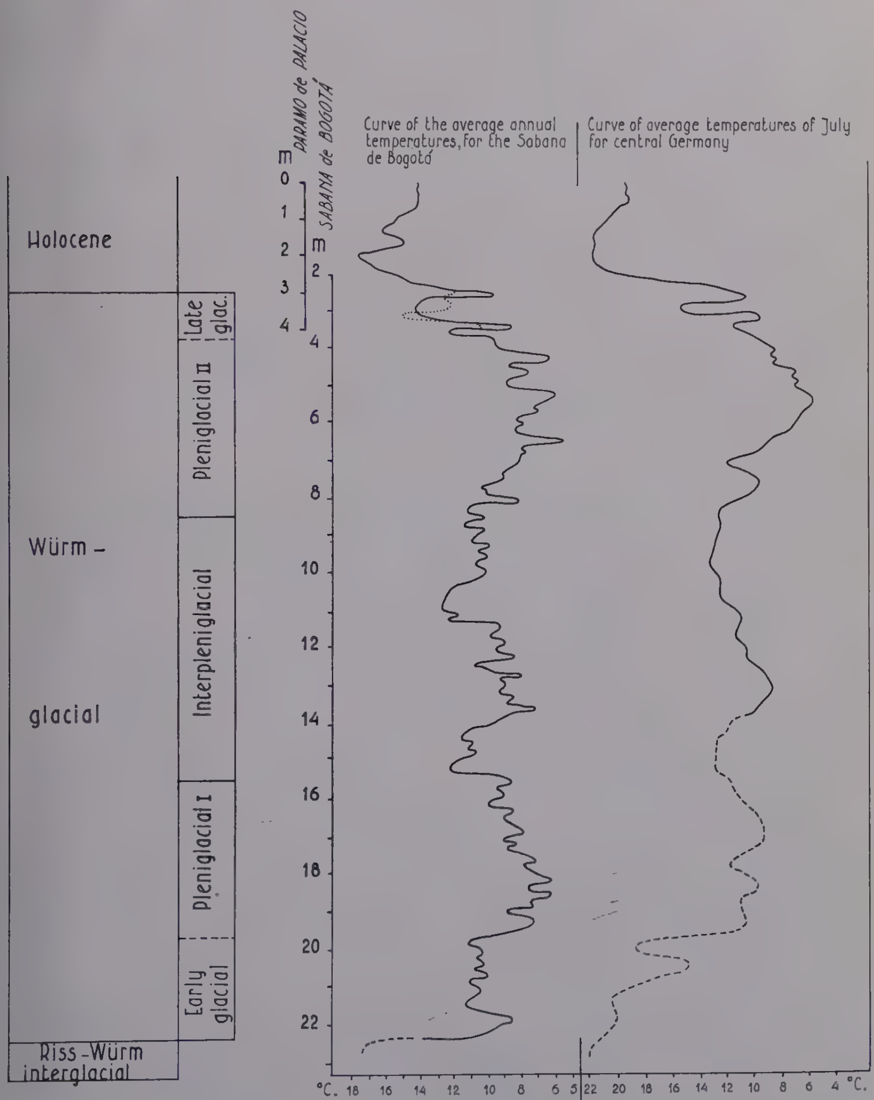


FIGURE 2. Temperature curve of the Upper Pleistocene and Holocene of the equatorial Andes (Sabana de Bogotá), compared with a curve for the Upper Pleistocene of Europe of H. Gross (1958).

(2) The glacials and interglacials, and also certain interstadials and stadials of the Colombian equatorial Andes, correspond exactly in time, in so far as the evidence goes, with those from other parts of the world. Our temperature curve corresponds remarkably well with Emiliani's and Gross's curves.

(3) Even the minor climatic fluctuations of the Holocene of the Colombian equatorial Andes correspond in time with the European.

Data From the Tropical Lowland

Sections from the lower Magdalena valley and from the Llanos Orientales (the plains with savannas and forests east of the Andes) are being analyzed on their pollen content, but the investigations are not yet finished and, at this moment, little more can be said than that important changes of vegetation have taken place in these regions since the final part of the last glaciation.

Other investigations are in progress on the coastal sediments of British Guiana. The first data have been published already (van der Hammen, 1959), and other papers are in preparation. Although C-14 data are not yet available, it seems possible to give at the present time an interpretation of the pollen diagrams.

FIGURE 3 is a simplified diagram of a 100-foot section of the coast near Georgetown, British Guiana. Three groups of vegetation are distinguished in this diagram: one comprising *Rhizophora* and other elements of the mangrove-swamp forest; the second, other forest elements; and the third, savanna elements.

In the interval between 73 and 32 feet the mangrove elements are completely dominating. From 32 feet to the surface we see a gradual fall of these elements, while forest and savanna elements increase. Doubtless this fall indicates a gradual retreat of the sea, and it is possible that the last sudden fall at 8 feet represents the influence of man (the construction of small dikes and similar structures). Between 94 and 75 feet a brown mottled layer is intercalated in the grey swamp clay. This layer contains in its upper part almost exclusively pollen of savanna elements and, in its lower part, half-mangrove elements, half-savanna, and forest elements.

From 94 feet to the base of the section at 101 feet, the mangrove-swamp elements dominate again.

It is quite clear that an important regression of the sea is reflected by the pollen content of the brown-mottled layer and, during the deposition of the upper part of the layer, the sea must have been far away (the few grains of *Rhizophora* found in this pure savanna layer are most probably reworked from older sediments, as most sediments in the coastal region contain abundant pollen grains of this mangrove genus).

The savanna layer seems to be lacking in the actual river valleys. Pollen analysis of series of sediments from the Demerara river near Mackenzie show that the same succession as in the upper part of our diagram is present there, but that the savanna layer and the underlying mangrove swamp clays are lacking. The base of the upper-mangrove swamp clays rests there immediately on the much older white-sand formation. This base lies at 125 feet in the Mackenzie section and at 75 feet, in the Georgetown section.

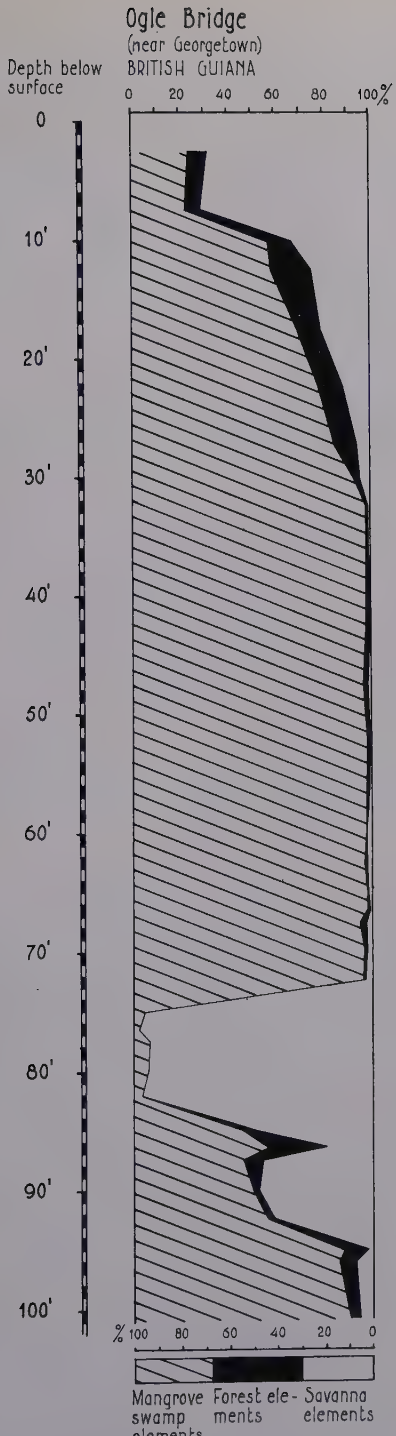


FIGURE 3. General pollen diagram of the upper part of the coastal sediments near Georgetown, British Guiana.

These data (together with similar still unpublished data from British Guiana) all point to the same phenomenon: an important regression of the sea accompanied by vertical erosion of the rivers, followed by an important transgression, first inundating the deeper river valleys and, after, the whole low coastal area. The upper swamp clays belong to the Holocene "Demerara clay" and, although C-14 dates are not yet available, it seems justifiable to correlate the important regression established at their base (or at the top of the "Coropina clay") with the regression of the Würm glaciation. We hope to have confirmation of this interpretation by means of radiocarbon analysis within a few months.

The interpretation of the facts given above implies that the mangrove swamps and forest of the coastal area of British Guiana, at some 7° N of the equator, disappeared during the last glacial and were replaced by grass savannas. This seems in contradiction with the fact we noted in the first part of this article that the last glacial in the Colombian Andes (at 5° north of the equator) was a pluvial period at the same time. This fact may be explained partly by the regression of the sea and the vertical erosion of the rivers, resulting most probably in a lowering of the ground-water table. However, if the climate had been pluvial without pronounced dry seasons it seems that on this clay soil a forest could have developed. There remains the possibility that the climate in the coastal region of British Guiana was pluvial (with a higher annual precipitation), but with one or two pronounced dry seasons during which the clay soil could desiccate so strongly that tree growth would be impossible for that reason. New and more abundant pollen-analytical data from the tropical lowlands probably will help in the future to solve this and similar important problems.

General Conclusions

We shall now consider for a moment the principal data obtained in northern South America, and compare them with the known published data from other parts of the world. Pollen-analytical data and C-14 dates from Europe, North America (including Hawaii), South America (including its equatorial northern part), Chile (Heusser, 1960), and Patagonia (Auer, 1958, and others) prove unequivocally that the climatic changes of the Upper Pleistocene and Holocene are contemporaneous, and that they belong to a world-wide phenomenon. The changes of humidity may be parallel or opposite, but the temperature changes are parallel. The maximum decrease of temperature during the last glaciation was approximately of the same order everywhere in the world (including the tropics) where it could be established. The temperature changes in general seem to be primary. Every theory that does not take into account these facts (especially the world-wide contemporaneity of changes of average annual temperature) and does not explain them satisfactorily cannot be the right one. It seems at this moment that only fluctuations of solar radiation (direct, indirect, or both) can give a satisfactory explanation of glacials, interglacials, and smaller climatic changes.

References

AUER, V. 1958. The Pleistocene of Fuego Patagonia. Part II: The history of the flora and vegetation. Annales academiae scientiarum fennicae. A, III. Vol. **50**.

Emiliani, C. 1955. Pleistocene temperatures. The Journal of Geology, Vol. **63**(6): 538–578.

Fairbridge, R. W. 1960. The changing level of the sea. Scientific American, Vol. **202**(5): 70–79.

Gross, H. 1958. Die bisherigen von C_{14} Messungen und Paläontologischen untersuchungen für die Gliederung und Chronologie des Jungpleistozäns in Mitteleuropa und der Nachbargebieten. Eiszeitalter und Gegenwart, Vol. **9**: 155–187.

Heusser, C. J. 1960. Late Pleistocene environments of the Laguna de San Rafael Area, Chile. The Geographical Review, Vol. **50**(**4**): 555–577.

Maarleveld, G. C. & T. van der Hammen. 1959. The correlation between Upper Pleistocene pluvial and glacial stages. Geologie & Mijnbouw, N. S. Vol. **21**(**2**): 40–45.

van der Hammen, T. 1959. First results of pollen analysis in British Guiana. Fifth Inter Guiana Geological Conference, Georgetown. Geol. Survey Dept.

van der Hammen, T. & E. Gonzalez. 1960*a*. Upper Pleistocene and Holocene climate and vegetation of the "Sabana de Bogotá" (Colombia, South America). Leidse Geologische Mededelingen, Vol. 25: 261–315.

van der Hammen, T. & E. Gonzalez. 1960*b*. Holocene and late-glacial climate and vegetation of Páramo de Palacio (Eastern Cordillera, Colombia, South America). Geologie & Mijnbouw, Vol. **39**(**12**): 737–745.

PALEOBOTANICAL RECORD OF SOLAR CHANGE

Herman F. Becker*
The New York Botanical Garden, Bronx, N. Y.

It is the primary task of the macropaleobotanist to study all available plant structures (leaves, flowers, bracts, fruits, and seeds), identify them, and place them into their proper taxonomic position. Ancient floras are thus reconstructed on a basis of progressive accuracy the younger they are in the geologic time scale. Certain basic observations are accepted as valid, and the findings are interpreted accordingly. The geologist assumes that the present is a key to the past, a maxim that is accepted by the zoologist and the botanist, and therefore the paleobotanist also avails himself of this prerogative. This implies a knowledge of the requirements for our regional or geographic floras: the tropical, the subtropical, the temperate, the arctic, and their intermediates, and also suggests immediately that the most recent or "modern" plants are by far better known than the most ancient. Paleozoic plants have left few, if any, recognizable descendants excepting the more encompassing taxa. Thus we distinguish Paleozoic members of the horsetails, the club mosses, ferns, and some conifers and cycads. It is only during the Mesozoic that the world flora begins to assume modern aspects. The Cretaceous represents a turning point, or the beginning of a more accurate knowledge and recognition of evolutionary lines leading gradually but inexorably to our Recent, living flora. As the gymnosperms are generically a relatively small group, it is the angiosperms that present the key to almost all of our modern taxonomic interpretation and to the radiating inferences regarding variability in individual plants and entire floras, evolutionary changes, migration, ecology topography, climates, and the movement of climatic belts through the ages. Since a great amount of data is obviously more conducive to exacting interpretation than spotty information, it is natural that we can draw only general conclusions regarding taxonomic status and climatological events for the earlier geologic periods. With the advent of an almost explosive evolution of the angiosperms during the middle-to-late Cretaceous, a much greater number of species and floras are observable than for any earlier period. This trend is highly progressive during the Tertiary, yielding a maximum of specific data for the Eocene, Oligocene, and Miocene.

Because of the relative slow and conservative evolutionary change in plants, index fossils on an epoch level are comparatively rare, but *Holmskioldia speirii* (Lesquereux) MacGinitie (*Porana speirii*) of the Verbenaceae is reliably indicative of the Oligocene. Invertebrate index fossils, however, are frequently employed and often indiscriminately relied upon when evidence to the contrary is available through paleobotany. It is also noteworthy that in the absence of animal fossils or geological evidence, plant remains may furnish the only clue to age and climatic determinations. Plant remains also provide an excellent tool to substantiate conclusions or correct them if they are erroneous or doubtful.

* Research Associate.

In fossil morphological structures we possess an immeasurable wealth of direct evidence for the reconstruction of a flora, and also a most accurate medium for their climatic interpretation. Many such structures display a similarity, even an identity to those of their living descendants, so that sometimes, by all evidence, a distinction between the fossil and the living species becomes problematical. In these cases it becomes self-evident that the fossils must be accorded climatic requirements identical to those of their living counterparts.

Sinnott and Bailey (1915) have determined that foliar morphology in families, genera, and even in species responds to prevailing climates, an observation that is successfully applied to fossil floras as well. Chaney and Sanborn (1933), Potbury (1935), MacGinitie (1937, 1941), Dorf (1942, 1955, 1959), and others have applied this method so successfully to climate determination in our western fossil record that it is now accepted as basic and almost infallible. Chief criteria for such determinations are (1) the size of leaves (over 10 cm., under 10 cm.); (2) organization (simple, compound); (3) margin (entire,

TABLE 1
ANALYSIS OF LEAF CHARACTERS IN PER CENT

	Organization		Length		Margin		Nervation	
	simple	comp'd	over 10 cm.	under 10 cm.	entire	non-entire	pinnate	palmate
Muir Woods, Calif.	77	23	27	73	23	77	77	23
Bridge Creek, Oreg.	80	20	27	73	15	85	70	30
Mormon Creek, Mont.	79	21	27	73	21	79	81	19
Range of per cent val.	3	3	0	0	8	8	4	11

nonentire, that is, lobed, scalloped, dentate, serrate); (4) nervation (pinnate, palmate); (5) dripping point; and (6) texture. To illuminate these climatically determined morphological differences, two contrasting, typical, living floras are chosen: (1) the tropical Barro Colorado assemblage from Panama, and (2) the temperate Muir Woods, Redwood Forest flora of California (Chaney and Sanborn, 1933). Any randomly chosen, contrasting, tropical, and temperate floras, however, will yield comparable results. The following table demonstrates the similarity of the most typically temperate Upper Oligocene Bridge Creek flora (Chaney and Sanborn, 1933) from Oregon with the living Muir Woods flora, and the recently described Lower Eocene Ruby Mormon Creek flora (Becker, 1960) from Montana, with both of the former (TABLE 1).

The small range of per cent values suggests an almost identical climatic requirement for these three floras. It is on the basis of such analyses that most of our western floras (FIGURE 1) acquired their proper climatic exponent. Accordingly, it is generally accepted that the Eocene epoch as a whole represented a tropical to subtropical time span with a temperature average near 70° F. The Eocene Mormon Creek flora of typical deciduous, warm-temperate leaf character, presents a morphological deviation and must have existed beyond the reach of the average Eocene tropical climate and precipitation,

perhaps at higher elevations or at the fringe of available moisture and temperature. These conclusions are based on the realization that not all plants are equally suited as climatic indicators, and that those with a known extensive latitudinal and longitudinal or altitudinal range must be omitted in the climatic interpretation of a flora. To this group belong many streamside elements such as willows, alders, and dogwood, but also pines and bracken ferns. If anomalous species are disregarded, the taxonomic status and established morphological features express accurately the prevailing climatic conditions of a given period. On the surface this procedure may seem an oversimplification of the correlation problem, and it is true that the additional factors of altitude and migration tend to influence the morphological expression of

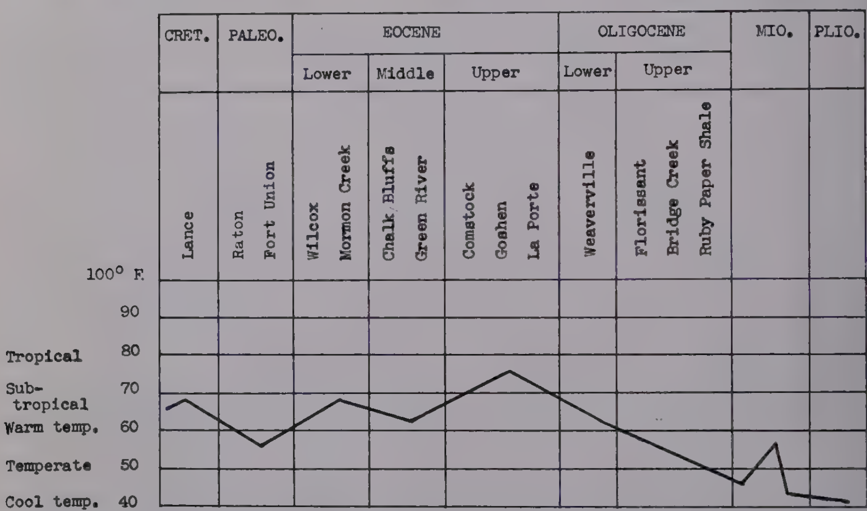


FIGURE 1. Generalized climatic regime of the Late Cretaceous and Tertiary in lat. 40° to 50° N of the western United States. Locations by state: Lance, Mont. and Wyo.; Raton, Colo.; Fort Union, Neb.; Wilcox, Ark.-Ky.-La.-Miss.-Tenn.-Texas; Mormon Creek, Mont.; Chalk Bluffs, Calif.; Green River, Colo.-Utah-Wyo.; Comstock, Ore.; Goshen, Ore.; La Porte, Calif.; Weaverville, Calif.; Florissant, Colo.; Bridge Creek, Ore.; Ruby Paper Shale, Mont. Adapted from Dorf, 1955.

otherwise homologous taxonomic units. Some species possess a wider range of adaptability than others and tend to merge peripherally with floras of different climatic requirements. These factors must be evaluated individually for each flora and considered in the over-all climatic aspects. In some respects the Mormon Creek flora displays its subtropical derivation as shown in the presence of *Bumelia*, *Cassia*, *Laurophyllum*, *Lonchocarpus*, *Nectandra*, *Persea*, and *Ptelea*.

Although the Late Oligocene Ruby Paper Shale flora lies within a mile of the Mormon Creek deposits, its taxonomic composition expresses the seasonal aspects of a dry-temperate association. It is primarily a deciduous-gymnospermous-herbaceous, strongly xeric flora with a complete lack of the subtropical element. Ecological habitats are clearly expressed in species that are conventionally referred to as aquatic, streamside, deciduous forest, coniferous forest, dry woodland, and xeric. The leaves are generally smaller than aver-

age, suggesting reduced precipitation. The floral composition is considerably more varied and modern than that of the Eocene flora, containing species of such genera as *Abies*, *Picea*, *Pinus*, *Acer*, *Betula*, *Carpinus*, *Celtis*, *Cercidiphyllum*, *Cercocarpus*, *Fagus*, *Fraxinus*, *Holmskioldia*, *Mahonia*, *Metasequoia*, *Populus*, *Ribes*, *Rosa*, *Sapindus*, *Sorbus*, *Ulmus*, and *Zelkova*.

In order to gain a regional climatic picture, all existing floras must be brought into climatic relation to each other resulting in a recognition of climatic belts and also in a recognition of prevailing topography and relief. A continued correlation of continental floras through subsequent epochs will reveal the shifting of these climatic belts. Paleobotanical evidence has demonstrated tropical or near-tropical periods in Alaska, Greenland, Spitsbergen and Siberia, and has traced their dispersal to present locations.

One of the most brilliant symposia (Ecological Monographs, 1947), entitled *Origin and Development of Natural Floristic Areas with Special Reference to North America*, elucidates the basic premises in aspects of paleobotanical ramifications, including paleoclimatology. Titles and participants are as follows: Just, Theodor, "Geology and Plant Distribution"; Chaney, Ralph W., "Tertiary Centers and Migration Routes"; Stebbins, G. L., Jr., "Evidence on Rates of Evolution from the Distribution of Existing and Fossil Plant Species"; Camp, W. H., "Distribution Patterns in Modern Plants and the Problems of Ancient Dispersals"; Cain, Stanley A., "Characteristics of Natural Areas and Factors in Their Development"; Mason, Herbert L., "Evolution of Certain Floristic Associations in Western North America"; Braun, Lucy E., "Development of the Deciduous Forests in Eastern North America"; and Raup, Hugh M., "Some Natural Floristic Areas in Boreal America."

Dorf (1959) has summarized the present status of paleobotanical climate determination and the shifting of climatic belts in his most excellent treatise entitled: *Climatic Changes of the Past and Present*.

References

Bailey, I. W. & E. W. Sinnott. 1915. A botanical index of Cretaceous and Tertiary climates. Science. **41:** 831–834.

Becker, H. F. 1960. The Tertiary Mormon Creek flora from the upper Ruby River basin in southwestern Montana. Palaeontographica B. **107**(4–6): 83–126.

Becker, H. F. 1961. Oligocene plants from the upper Ruby River basin, southwestern Montana. Geol. Soc. Am. Mem. **82:** 128 pp.

Chaney, R. W. & E. I. Sanborn. 1933. The Goshen flora of west central Oregon. Carnegie Inst. Wash. Publ. **439:** 103 pp.

Dorf, E. 1942. Upper Cretaceous floras of the Rocky Mountain region. Carnegie Inst. Wash. Publ. **508:** 168 pp.

Dorf, E. 1955. Plants and the geologic time scale. Geol. Soc. Am. Spec. Paper. **62:** 557–592.

Dorf, E. 1959. Climatic changes of the past and present. Contrib. Museum Paleontol. Univ. Mich. **13**(6): 181–210.

Ecological Monographs. 1947. Symposium. Ecol. Monogr. **17**(2): 125–234.

MacGinitie, H. D. 1937. The flora of the Weaverville beds of Trinity County, California. Carnegie Inst. Wash. Publ. **465**(3): 83–156.

MacGinitie, H. D. 1941. A middle Eocene flora from the central Sierra Nevada. Carnegie Inst. Wash. Publ. **534:** 178 pp.

Potbury, S. S. 1935. The LaPorte flora of Plumas County, California. Carnegie Inst. Wash. Publ. **465:** 29–81.

Sinnott, E. W. & I. W. Bailey. 1915. Investigations on the phylogeny of angiosperms; foliar evidence as to the ancestry and early climatic environment of the angiosperms. Am. J. Botany. **2**.

NEW HORIZONS FOR THE ATMOSPHERIC SCIENCES

Alan T. Waterman
National Science Foundation, Washington, D.C.

I thought it might be of interest if I were to review briefly the major circumstances influencing the progress of the atmospheric sciences in the last decade and some of the plans and prospects for the future.

Scientific research has accelerated sharply in nearly all fields in recent years. Meteorology, however, has certain characteristics that, although not completely unique, play a much larger role in relation to its progress than they do in other fields. One is its dimensions—since the whole of the earth's atmosphere is literally its laboratory—and the other is its special international character, which it shares with much of geophysics. Although most scientific research progresses with little regard for political and geographical boundaries, much of meteorological research is absolutely dependent upon world-wide cooperation. For these reasons the International Geophysical Year, the Antarctic program, and the space research program have all served to stimulate additional interest and activity in meteorology and related sciences.

The International Geophysical Year (IGY) of 1957–1958 and the intensive work that was done in all the fields of geophysics produced vast quantities of meteorological data and introduced new and spectacular instruments and techniques for the acquisition of such data. This effort, based on the contributions of so many nations, had its roots, of course, in the first and second polar years of 1882 and 1932, which indicates that the meteorologists of the world have long been working together in an organized way to develop the tools and enlist the cooperation needed for the acquisition of synoptic data so necessary to progress in their field. During the Second Polar Year the significant new instrument was the radiosonde. The IGY made notable use of rockets and witnessed the launching of the first earth satellites. The earlier satellites, you will recall, transmitted a variety of data related to the phenomena of the upper atmosphere but were not, strictly speaking, the so-called weather satellites.

Back in 1951 (six years before the first earth satellite was launched), S. M. Greenfield and W. W. Kellogg of the Rand Corporation prepared a study on the *Feasibility of Weather Reconnaissance from a Satellite Vehicle.*

In 1957 Harry Wexler,[1] of the United States Weather Bureau, Washington, D.C., describing an "ideal" weather satellite, commented that this would not, of course, be available for some years. In 1960, much earlier than expected, the United States launched not one but two weather satellites: TIROS I and II which, Wexler declares, come pretty close to the ideal he described.

During its two months of active life, TIROS I was producing photographs at the prodigal rate with which the Sorcerer's Apprentice produced brooms and pails! More than 20,000 photographs of the earth's cloud covering were obtained, many of which were of much higher quality than meteorologists had dared to hope. The photographs served a dual purpose. Some of them were used immediately as a basis for forecasting, and the most interesting ones were set aside for research and analysis.

About 300 weather maps synthesizing thousands of TIROS cloud photographs showing schematic cloud distribution and organization have been sent to forecasters for appraisal of their value in weather prediction. The cloud analysis charts are now being sent over the Weather Bureau's weather facsimile network to about 650 government, military, and commercial receiving stations in 330 cities in the United States and Canada. In addition, the cloud analyses charts based on the TIROS photographs are being transmitted over teletype and by radio facsimile to other countries and have been furnished for support of the annual summer resupply mission to Antarctica.

Although the wide-angle camera on TIROS II failed to come up to expectations, the performance of the narrow-angle camera was, if anything, better than the corresponding camera on TIROS I. More significantly, however, TIROS II is equipped with five different infrared sensors measuring various components of radiation from the earth and sun. It is anticipated that the information thus received will give us a firm grasp on the whole problem of energy input and output.

The next step, for which plans are already under way, is the development of a common system of meteorological observation satellites.

The space program reacts in an interesting way with the atmospheric sciences. The rockets and satellites provided by the space program constitute important new instruments for research in the upper atmosphere. At the same time, the space program is itself directly dependent upon research in the atmospheric sciences. All sorts of information is needed with respect to the nature and composition of the atmosphere at the higher levels in which space craft will be operating. The discovery of the Van Allen radiation belt opened up a whole series of questions as to its nature and gave renewed vigor and direction to upper atmosphere research and especially to studies of solar-atmosphere relationships. Other questions relating to the density of the atmosphere have a bearing on the important re-entry problem. Research related to these problems is, of course, supported directly by the National Aeronautics and Space Administration, Washington, D.C., which maintains close liaison with other agencies supporting research in the atmospheric sciences.

Another event that has had a notable impact on the atmospheric sciences is the program being carried on in the Antarctic. Begun under the IGY, Antarctic research embraces a wide variety of meteorological studies as well as research in other fields. In addition to the meteorological research being carried on by the nations who participated in the Antarctic program, the International Antarctic Analysis Center at Melbourne is studying synoptic weather patterns. Data have now been collected by individual weather stations in the Antarctic for a sufficient period to enable meteorologists to see cycles and patterns. The ultimate objective is to establish relationships between synoptic situations in Antarctica and in the rest of the world.

An enumeration of major factors influencing the development of atmospheric research in the last decade would certainly include the efforts that have been devoted to weather modification. The rash of attempted commercial rainmaking activities that sprang up as a result of the intriguing discovery of Vincent Schaefer[2] and the exploratory cloud-seeding experiments sponsored by

the Government in the late 1940s and early 1950s prompted Congress[3] to create an Advisory Committee on Weather Control in 1953 "to study and evaluate public and private experiments on weather modification." The work of this committee, as well as studies made by the Council of the American Meteorological Society[4] and the Committee on Meteorology of the National Academy of Sciences,[5] pointed to the need for much more fundamental knowledge of cloud physics and other meteorological phenomena as basic to possible weather modification and control.

As a matter of fact, modifications of the Langmuir-Schaefer cloud-seeding techniques, although studied in carefully planned field experiments and to some extent in the laboratory, have thus far given no evidence of highly significant increases in precipitation. Most such experiments have come out with negative results when they were planned and analyzed by sound statistical methods. It became clear, therefore, that much money and effort would be wasted unless and until more accurate scientific knowledge and understanding of the actual precipitation process could be obtained. This means careful basic research by highly competent individuals into the nature of the precipitation process both in the laboratory and in the natural environment. Thorough study, too, must be made of atmospheric conditions that may be amenable. Incidentally, from the standpoint of practical use, there are obviously two problems: the increased production of precipitation and, in certain circumstances such as severe storms, its prevention.

We should not forget that from an over-all scientific point of view, natural processes of precipitation, both in general and in local storms, involve energies of order of magnitude far greater than any that man can produce even with present thermonuclear reactions. Therefore the problem consists of looking for synoptic situations where nature's forces are in balance, so to speak, or in unstable equilibrium. If such situations can be pinpointed in time and location, then, and only then, would it seem possible to introduce a man-made source of energy or some catalytic agent that would affect the natural phenomenon to any appreciable extent. For example, if we find a cyclonic disturbance in the act of formation, it may be possible before the momentum of the process has become large to aid or hinder the natural process.

In physics there are recognized general methods for greatly increasing an effect occurring in a physical system. The first is usually designated as the regenerative process and is illustrated by the principles involved in electronic magnification of faint radio signals where a mechanism is set up with a continual feed-back into the original system that causes the effect to grow immeasurably. This is also exemplified in the liquefaction of air where part of the air that is cooled is used to cool the incoming air. A second method is the initiation of what is popularly called a chain reaction, and the production of atomic energy is a familiar example. A third utilizes the principle of "resonance," wherein the amplitude of a periodic phenomenon can be increased enormously by a small, properly timed impulse, as in the familiar example of pushing a child in a swing.

As a result of the recommendations of the Advisory Committee on Weather Control,[6] of which the late Howard T. Orville was chairman, Congress passed

Public Law 85-510 of July 11, 1958, directing the National Science Foundation "to initiate and support a program of study, research, and evaluation in the field of weather modification." The foundation appointed an advisory panel in this field and began to expand its program of grants and contracts for basic research in weather modification as well as its evaluation efforts. Since the law was passed, the foundation has obligated close to $3 million for research and evaluation on weather modification.

To recapitulate: a number of factors have influenced the growth of the atmospheric sciences in the last decade including, most notably, the International Geophysical Year and the Antarctic programs, the development of rockets and satellites as research tools, the need for new knowledge to implement the exploration of outer space, and the demand for a sound scientific base for weather modification. Nor should I neglect to mention the brilliant work of John von Neumann[7] in the application of the electronic digital computer to the solution of meteorological forecasting problems, thus making it possible to take advantage of the huge capacity and speed on the part of these machines for the solution of many highly complicated equations involving huge quantities of data.

Let us turn now to some of the problems. Until recently, research in meteorology and related fields has suffered from lack of both funds and manpower, to say nothing of public understanding of the need. In the foundation we have consistently supported basic research in the atmospheric sciences and, in 1958, we gave recognition to the growing importance of the field by establishing an Atmospheric Sciences Program within the Division of Mathematical, Physical, and Engineering Sciences. The new program was set up to serve as a point of focus for needs in the field and to provide the long-term, stable support heretofore lacking for research in this area.

The scientific community performed a notable service on behalf of the atmospheric sciences by taking energetic action to call attention to the need and to prescribe remedial action. In 1956 the National Academy of Sciences established the Committee on Meteorology ". . . for the purpose of bringing together scientists from meteorology and related physical and geophysical fields to view in broad perspective the present position and for the requirements of meteorological research and to recommend the general outlining of a program which would accelerate progress in this important field."

After two years of intensive work, the committee issued, in January of 1958, a significant report entitled *Research and Education in Meteorology*,[8] which recommended a number of steps that needed to be taken to insure our national strength in this important field. These related to three principal points: greatly increased support for basic research in meteorology at the universities; the establishment of a national institute of atmospheric research; and an intensive effort to meet the manpower needs in the field, both through improved professional opportunities and through revision and expansion of university meteorological curricula.

It is pleasant to report that in a city where significant reports by competent people all too often gather dust on the shelves, the report of the Academy's Committee on Meteorology has resulted in prompt and effective action. Just

five days after the Committee's report appeared, representatives of 11 institutions, meeting in New York, N.Y., on January 31, 1958, approved a resolution endorsing the recommendations of the Committee. One month later the group met at the University of California in Los Angeles and constituted itself the University Committee on Atmospheric Research.

The new group, now familiar to everyone as UCAR, went to work at once to implement the Committee's recommendations and, specifically, to formulate a research program for a national institute of atmospheric research and to plan the physical plant, equipment, and personnel that would be needed to launch such a program.

In March 1959, the University Corporation for Atmospheric Research was constituted and, in June 1960, the foundation had the pleasure of announcing a grant to the corporation that would make it possible to begin the work of establishing a National Center for Atmospheric Research, under the direction of Walter Orr Roberts. The function of the center is to enhance the effectiveness of atmospheric research by fostering a vigorous interdisciplinary effort and providing the large facilities and technological support that are required in order to make significant advances in the atmospheric sciences. The center will supplement and not supplant the research effort of the universities and will extend their efforts by providing facilities that otherwise would be unavailable or too extensive to be managed by individual institutions.

At the present time Roberts expects the research problems to fall into four main categories: (1) atmospheric motion; (2) energy exchange processes in the atmosphere; (3) water substance in the atmosphere; and (4) physical phenomena in the atmosphere. The staff will represent many different disciplines involved in the atmospheric sciences, including physics, chemistry, engineering, mathematics, and meteorology.

In November of last year, the foundation and UCAR announced the selection of Table Mountain, near Boulder, Colo., as the site for the new center. Thus within three years, almost to the day, of the Committee of Meteorology's recommendations, effective steps have been taken toward carrying them out. This includes not only the establishment of the center, but also constructive work by the American Meteorological Society and others towards meeting some of the pressing problems relating to manpower and education in the field.

The American Meteorological Society, with support from the National Science Foundation, has undertaken to acquaint students with the career possibilities in the field of meteorology. Under the Visiting Lecturer Program, research meteorologists visit numerous colleges and universities throughout the United States "(1) to demonstrate the vitality and importance of the atmospheric sciences; (2) to present one or more of the challenging problems of the atmosphere; and (3) to make known the opportunities available to graduate students and research scientists."*

The society also sponsored a summer atmospheric sciences program for secondary school students at the Loomis school in Connecticut, under the direction of Vincent J. Schaefer. Work shops and undergraduate research participation programs have been made available for the purpose of acquainting

* *Meteorology on the Move*, National Academy of Sciences—National Research Council, Publication 794, 1960.

undergraduate science students with the problems and techniques of meteorological research. Both the Alfred P. Sloan Foundation and the Ford Foundation both in New York, N.Y., have made funds available for fellowships in the field and, of course, NSF fellowships in a number of categories are available for qualified students, teachers, and research leaders in meteorology.

The National Academy of Sciences' Committee on Atmospheric Sciences, in its most recent report,* points out that whereas the average number of meteorology students at the doctorate level during the 10-year period 1950 to 1959 has been 12 per year, the universities should produce about 50 to 60 per year to provide the required manpower needs to support the recommended level of research activity at the universities and the proposed Center for Atmospheric Research.

UCAR studies indicate that this need can be met without major additions to existing staff or facilities. The crux of the matter is attracting capable students to the field, and to this problem there must be continuing attention.

Another point that should be noted is that research in meteorology and research in oceanography overlap at many points. Oceanography is another of the large critical areas of science that, it is realized, need a greatly intensified effort on the part of scientists in this country. It, too, has been the subject of special study by a National Academy—Research Council Committee on Oceanography, and a program has been presented to the government recommending the steps that should be taken to assure adequate progress in that field also. It is of interest to note that the Committee on Atmospheric Sciences and the Committee on Oceanography, recognizing areas of common interest, have "jointly organized a panel on air-sea interaction, with membership comprising nine outstanding scientists drawn from the two disciplines having prime interest and concerns in the problems common to both."

Here I should perhaps inject a note of caution. Gratifying progress has been made recently in recognizing and meeting the needs of the atmospheric sciences and oceanography, but these are only two of a number of critical areas that must compete for funds and manpower. Most of these areas of research require large and expensive equipment. Our problem in Washington, D.C., is to allocate available federal funds among these fields in a way that will enable all of them to advance significantly. Unfortunately, this means that no one of these fields is going to acquire all that it believes it needs, or will have its requirements met as early as would seem desirable. Nevertheless we should be encouraged that recognition has been accorded these special fields—several were specifically cited in the presidential budget message last week—and that programs have been established that will provide a basis for future growth and expansion.

It might be pertinent to recall that at the time of my first appearance before the American Meteorological Society nine years ago, the Foundation itself was operating under an appropriation of $3.5 million. Our appropriation for the current fiscal year is slightly under $176 million, but if the needs of basic science and science education are to be met adequately, our appropriation must continue to grow substantially in the future.

In the early years of the foundation, the limited funds available restricted our

* Ibid, p. 10.

activities pretty specifically to the support of individual research projects and to the award of fellowships in the sciences. As funds have been increased, we have been able to expand our activities to include some of the other pressing needs of science, such as funds for facilities and equipment, for improving science teaching in the secondary schools, for science course-content improvement, and for laboratory construction and renovation. We have also been able to increase and extend research support to include other types of grants, such as, for example, broader grants to cover coherent areas of research.

We have made several grants of this type in the atmospheric sciences. At Harvard University, Cambridge, Mass., the foundation is supporting a program of atmospheric studies in the general area of physics, applied physics, and applied mathematics. The purpose of the program is to build a small competent group of workers to engage in aspects of atmospheric study that can be advantageously treated by deductive scientific methods. The ultimate hope is that many students trained in the standard physical sciences will enter the field of meteorology regularly. At the University of Chicago, Urbana, Ill., the presence of a group of cloud physicists has made it possible to establish a program of cloud-physics research dealing with the water resources of clouds. The research covers all the factors believed to be important in precipitation mechanisms, and the grant was awarded for a three-year period.

As the university research departments are taxed to an ever-growing extent for the conduct of research of particular interest to government or industry, it becomes more and more important for them to have uncommitted funds with which to carry out their own plans and strengthen their own research groups. Both the National Science Foundation and the National Institutes of Health of the Public Health Service, Bethesda, Md., have recently inaugurated a type of institutional grants program in which modest sums are made available to the universities, with no restrictions save that they be used for the advancement of science. The latest report of the President's Science Advisory Committee, "Scientific Progress, the Universities, and the Federal Government," emphasizes the integral relationship of basic research and graduate education and calls for increased federal support to enable the universities to meet the needs of the nation's research effort.

Last year the foundation took the initiative in the establishment of an Interdepartmental Committee on the Atmospheric Sciences, comprising representatives of those departments and agencies having research interests in these fields. Surprisingly perhaps, the list is rather long. It includes the Departments of Defense, Agriculture, Interior, and Commerce (Weather Bureau), the Atomic Energy Commission, the National Aeronautics and Space Administration, Federal Aviation Agency, the Department of Health, Education, and Welfare, and the National Science Foundation.

During fiscal year 1960 the federal government expended $54.1 million for the support of basic and applied research in the atmospheric sciences and, in fiscal 1961, the figure is expected to reach $67.4 million. These sums include expenditures for rockets and satellites employed for meteorological purposes. The Interdepartmental Committee, which now operates under the Federal Council on Science and Technology, affords each agency opportunity to learn

what the other agencies are doing in the field of atmospheric research and thus fosters effective coordination and helps to prevent overlap and duplication. The committee also advises the council regarding measures that it feels should be taken to advance the progress of atmospheric research in the United States.

Last year a special committee, appointed by the National Academy of Sciences at the request of the Secretary of Commerce, reported on *The Role of the Department of Commerce in Science and Technology*.[9] The committee, headed by Mervin J. Kelly of the Bell Telephone Laboratories, Inc., urged that the United States Weather Bureau be strengthened and expanded, particularly in its research functions. The committee felt that the research of the Weather Bureau should be geared to the rate of growth of the National Center for Atmospheric Research. For years the Weather Bureau has been forced to operate with insufficient funds and has suffered from a lack of understanding of the needs for research to backstop and aid its operational activities. The Academy's Committee on Atmospheric Sciences strongly endorsed the recommendations of the Kelly committee with respect to the Weather Bureau, and it is hoped that these will receive the favorable attention of those who will be responsible for guiding the Department of Commerce in the next four years.

The opportunities for progress in atmospheric research are now very great, far greater than at any time in the past. Not only are our instruments and techniques for the study of atmospheric phenomena vastly improved in accuracy and variety, but now that almost limitless potentialities exist for experimental observations through the limits of our atmosphere into outer space, the study has become a thoroughgoing three-dimensional one. This is, of course, what it has always been, but until recently the limitations upon our ability to observe have confined it essentially to two dimensions. At the same time, the increasing work in oceanography has shown the importance of linking this subject continuously with phenomena in the atmosphere above the seas. However, of much greater significance is the opportunity we now have to study the really fundamental problem, which is not merely the local circulation of the atmosphere of the earth and its associated climatic and weather fluctuations, but the over-all view of the relationships between the sun and the earth as a gigantic thermodynamic system.

The IGY observations showed clearly for the first time that the earth must be regarded as traveling in a real sense through the outer atmosphere of the sun. We have known in the past, of course, that disturbances on the sun affect our earth, but we now realize that the phenomena that produce these more striking fluctuations are actually a continuous component of our own physical system. The whole might be regarded as a form of heat engine with the sun as source and the earth as receiver. We therefore need to study much more carefully the output of radiation from the sun in its enormous range from cosmic radiation throughout the electromagnetic spectrum to the long-wave heat radiation. We are also interested in the degree to which this incident radiation is reflected by the earth, what is absorbed and where, and what is reradiated. This is indeed a cosmic problem and one that should be fascinating to investigators in a great variety of the sciences. In the coming years we shall have ample opportunities for experimental observations of the sun's

true output, including fluctuations, and of the effects that are produced in our atmosphere and upon the earth's surface. When one considers the speed and facility with which earth satellites can provide us with data, it is evident that a whole new host of investigators from many scientific disciplines should be brought in and the work systematized in order that analyses can be made and conclusions drawn. This is surely one of the grandest and most impressive scientific problems that man has been privileged to undertake.

At the same time, however, we must realize that if in the course of this global study we should acquire the ability to alter the processes of nature on a macroscopic scale, we should do well to enter upon such a course with caution. We are here dealing with forces almost beyond our imagination, and we must not forget that we could unwittingly touch off a catastrophic reaction. For this reason, it is more than ever important that these subjects be approached through very careful and thorough basic research in order to guard as best we can against the possibility of losing the control that we may seek.

It may be seen from even a cursory view of the current status of research in the atmospheric sciences that here we have at once a problem of tremendous dimensions and a marvelous opportunity. Both from the scientific and the practical point of view the potentialities of the field are such that the best thinking and collaboration of scientists from many disciplines will be required to do the subject justice.

References

1. Wexler, H. 1957. The satellite and meteorology. J. Astronautics. **4**(1): 1–6.
2. Schaeffer, V. J. 1946. The production of ice crystals in a cloud of supercooled water droplets. Science. **104**: 457–459.
3. United States Congress. 1953. Section 10(a) of the Act of August 13. As amended, Sec. 311.
4. Bulletin of the American Meteorological Society. 1957. Statement on weather modification. **38**(6): 366.
5. Research and Education in Meteorology. 1958. An interim rept. of the Committee on Meteorology to the NAS-NRC. Washington, D.C.
6. Advisory Committee on Weather Control. 1957. Final Report. 2 vols. Washington, D.C.
7. Charney, J. G., R. Fjortoft & J. von Neumann. 1950. Numerical integration of the barotropic vorticity equation. Tellus. **2**(4): 237–254.
8. *Ibid.*
9. National Academy of Sciences. 1960. The role of the Department of Commerce in science and technology—a report to the Secretary of Commerce by a special advisory committee of National Academy of Sciences.

Part VIII. Climatic Change and Man

EVIDENCES OF CLIMATIC FLUCTUATIONS IN SOUTHWESTERN PREHISTORY*

Terah L. Smiley
Geochronology Laboratories, University of Arizona, Tuscon, Ariz.

Introduction

Weather and climate are probably the most ubiquitous topics of conversation, and they enter into almost every piece of writing in the physical and biological sciences dealing with historical problems. In spite of all that is said and written we have comparatively little knowledge on the subject. Most of our information on past climates and weather conditions is of a qualitative nature; for example, it was cooler or warmer, it was drier or wetter. Without instruments to measure the intensity accurately, people seldom have the same relative response to a particular weather condition: each has his own "comfort index," which may or may not agree with others living in the same area.

In working with historical documentation that lacks instrumental measurements, it is difficult to evaluate the records properly. If, however, an author relies on the presence or absence of certain plant and animal life forms or the changes in population of those forms, a more reliable estimate can sometimes be given to such documents. Without weather records based on instrumental measurements, students of past weather and climate have had to turn to the physical and biological sciences for aid through the study of the physical and chemical properties of the sediments and of the fossil plant and animal remains.

Here, however, we are in difficulty because this leads to problems in ecology, particularly in the range of a tolerance a plant or animal has in regard to climatic or weather conditions. Few ecologists agree on the range of such tolerances until the absolute limits are reached. In a sense, plants and animals have a greater tolerance for changes in moisture than they do in temperature, but even in the latter there are wide variations. For example, certain plants can survive for long periods of time with the temperature slightly above the freezing point of the liquids in its cells, but if the temperature drops far below that point even for one night, the cells rupture and kill the plant. It is the extreme drop or change that causes the damage. These tolerances must be carefully evaluated before we place much reliance on their ability to serve as indicators. Another problem of importance concerns soil moisture. If a particular vegetative group is growing in an area where soil moisture is at the lowest amount the plants can tolerate, a slight drop will cause them to die and be replaced by other plants not requiring as much moisture. The amount of moisture can be greatly raised under these conditions, however, and there will be little discernable change.

Because of these uncertainties in range of tolerance conditions, our interpreta-

* Contribution No. 44: Program in Geochronology, University of Arizona, Tuscon, Ariz.
Contribution No. 31 from the University of Arizona Arid Lands Program, supported in part by The Rockefeller Foundation, New York, N.Y.

tion of past climate is based on a study of the plant and animal life as well as on a study of the chemistry and physical aspects of the deposits themselves. The micro- and megafossil remains are evaluated in light of the physical and chemical properties of the deposits, and it is believed that a more accurate picture of paleoclimates will result from this integrated type of approach.

I define climate as a composite, or aggregate, of the day-by-day weather conditions expressed over a "long" period of time; weather simply as the conditions of the atmosphere at a given time and place. It has been only within the last decade that accurate measurements of the elements have been available and that a proper picture of world climates can now be determined. Before this only sweeping generalizations could be made based on such factors as vegetation, relative temperature, and amounts of moisture, or other single types of indicators. Numerous systems have been devised such as those employed by Köppen (1936), Hettner (1930), de Martonne (1948), and Thornthwaite (1933 and 1948). These classifications can be used as long as one is not seeking detailed knowledge of quantitative data. If we can do no better than this with modern climates, what can we expect to do with past climates on which we have very little quantitative information?

As is well known, climate is not static; rather it is constantly undergoing certain alterations, and the various "averages" of the elements that make up weather and climate are constantly undergoing changes. Very little is understood of climatic change, however; in fact, so little is known on climatic change that in October 1962, the Advisory Committee on Arid Zone Research, United Nations Economic and Social Council, and the World Meteorological Organization will hold an international symposium on this general topic especially in regard to arid areas. I have never seen in print any definition of "climatic change," and I blame no one for not attempting to say what is meant by this term. In the semiarid southwestern part of the United States, the annual rainfall varies by as much as 40 per cent from one year to the next. With such a high variable as this, it is most difficult to determine if this area is undergoing any "change," especially since the records extend only one-half century back. Statistically speaking, there has been a general drop in yearly amounts for about the past 15 years (Von Eschen, 1958).

If we use plants and animals as climatic indicators we must also keep in mind that they do not measure meterological amounts of rainfall; rather they are a measure of the amount that penetrates the ground, forms lakes and cienegas or, in some manner, becomes usable for plants and subsequently animals. The plants must have soil moisture, and the type of plant cover is a direct result of the amount that can be effectively used. Herbivorous animals live off the plants: browsing animals requiring one type of plant cover and grazing animals needing another type of plant cover. Carnivorous animals live off the herbivorous types resulting in a rough ratio between the two types present in any given area. These animals have a wide range of feeding grounds and they move over the ground rather freely. Because of their mobility they can endure harsh environments for short periods of time even though they would prefer more "normal" conditions. Plants can also migrate under favorable conditions and remain in isolated areas favorable to

their needs even though the surrounding area is totally unfit for their regeneration.

Much has been written and said in the use of soils as climatic indicators. Soil moisture and temperature play an important role in soil chemical conditions, the latter controlling ion availability and uptake in plants. Recent research on trace elements in plant remains (Martin *et al.* 1960) using emission spectrographic techniques indicate that certain elements in plants such as nitrogen and manganese are a function of climatic temperatures.

The basis of the geochemical facies principle (Damon *et al.* 1960) is that the solubility and oxidation states of many elements are highly dependent on oxidizing (Eh) and alkalinity (*p*H) conditions of the soil. The Eh and *p*H are in turn influenced by the environment and climate of the deposits. Aqueous environments are limited at high Eh by decomposition with a release of oxygen and at a low Eh by the reduction of hydrogen ions with release of hydrogen gas. Arid environments tend to produce high *p*H and high Eh conditions resulting from them (for example, the oxidation of ferrous and manganous compounds, the oxidizing of sulfides to sulphates, the precipitation of carbonates, and the burning of organic matter). Humid conditions tend to form a reversal of these processes. In short, the composition of sediments, soils, and the trace element content of organic matter varies according to climatic conditions.

Erosion and sedimentation are examples of geologic processes that are common today and, certainly, evidences of past erosion and sedimentation are found in rocks representing the geologic past of this planet. A detailed study of the types of sediments, the form the sediments have taken since deposition, and the plant and animal materials contained in those same sediments will yield definite information on climate and probably climatic change. Dating these sediments gives the time periods when certain physical processes were in operation.

The foregoing are but a few examples of several methods now being employed to determine past climatic conditions. The accuracy with which these conditions can be determined depends largely on the number of techniques employed and how well these various techniques can be integrated into a composite interpretation.

The Geographic Setting

The present geographic relation of the Southwest to the remainder of the North American continental land mass and the present-day atmospheric circulation patterns cause it to be semiarid to arid. It has been semiarid or arid for as long as these conditions have been the same. Generally speaking, the Basin and Range Province, of which the Southwest is a part, came into existence approximately 25,000,000 years ago during the Miocene epoch. The present land forms have probably been the same since the Upper Pliocene times. Local volcanism, uplift, erosion, and deposition have caused minor changes in topography, however. The latest volcanic activity known has been dated at about 1100 A.D. Certain evidences of present tectonic activity are

to be found in contemporary fault zones where large slip faults in sedimentary deposits are active.

Scattered throughout the Basin and Range Province are numerous playa lake areas, some of which have remnants of well defined beaches and terraces. At the present time these playa areas are covered with shallow water after heavy rains or after snows have melted in the nearby mountains. These playas indicate more effective moisture conditions in the past; the dating of the beaches and terraces tell us when in the past the areas were filled with water. The fresh appearance of many of the playas indicate that the period was not too long ago: sometime within about the last 10 millennia; otherwise erosion would have destroyed more of these soft sediments.

Postglacial Climate

Exposures made by arroyo cutting, road and railroad cuts, and well cores indicate that in the middle reaches of the streams in this area postglacial deposits generally are unconformably resting on an older "basal" clay material. In several localities this clay contains animal remains of the Blancan fauna that lived during the late Pliocene and early Pleistocene epochs. Widespread sheet erosion (planation) must have been prevalent at this time because nowhere has as yet been found an indication of a soil horizon on the clay. It is impossible to determine how long erosion had been in progress; all we know is that it stripped away all overlying materials.

Man was living here at this time, and he may well have been here earlier but if so the erosion also carried away or otherwise destroyed all evidence of his presence. On the surface of the older clay are found man-made fireplaces containing charcoal fragments and other remains of man such as points, scrapers, and choppers (Haury *et al.*, 1959; Antevs, 1959; Lance, 1959). There are also bones and teeth of animals hunted and killed by man including the mammoth (*Archidiskodon*), tapir (*Tapirus*), horse (*Equus*), camel (*Calelops*), sloth (*Northrotherium*), bison (*Bison*), and carnivores such as the dire wolf (*Canis dirus*), and large cats (*Felix atrox*) (Smiley, 1959). Sediments of varying thickness covered and otherwise buried these remains in the intermontain valleys and on the flood plains.

The change of physical conditions that caused a shift from erosion to sedimentation in the middle reaches of these streams occurred about 11,000 to 12,000 years ago. This could have been brought about by a shift in land elevation or a change in climate; the evidence indicates that it was the latter. Indications of a short but intense "dry" period during this general time have been found by Broecker and Orr (1958) in the Lake Bonneville and Lake Lahontan areas of the Great Basin Province, and by Clisby and Sears (1956) in the San Augustin plains of the Basin and Range Province. It could have been that this intensive "drought" caused widespread destruction of vegetation in the highlands and, coupled with a change in storm patterns, allowed the deposition of eroded sediments in the middle reaches of these same streams.

Remains of the animal population existing at the time that have been found and studied are of little climatic value because these same animals had the ability to move great distances while seeking food. Nevertheless, all of the

large animals listed were soon to become extinct for one reason or another. Only the deer, bear, and other similar large animals survived.

The charcoal fragments from the fireplaces and the pollen grains from the sediments deposited during this time indicate that the plant species were essentially the same then as they are now. There undoubtedly were differences, however, in the percentage of species present. A fragment of charcoal and several pollen grains of hickory (*Carya*) from the sediments represents the only identified species found that is not growing in the area today.

A tentative conclusion of the foregoing evidence is that at that time the climate was much the same as it is today, perhaps a bit wetter, as indicated by the presence of hickory.

Water may still have been present in the Willcox playa (glacial Lake Cochise) of southeastern Arizona through this period of time, as artifacts and firepit material of this general period, from the Sulphur Spring stage of the Cochise culture to very recent pottery-making peoples have been found along the lowest beach line associated with this lake (E. B. Sayles, personal communication). The effective precipitation was decreasing, and the lake gradually dried. Large sand dunes were formed on the northeast side of the dry lake bed and in other areas of the Sulphur Spring Valley to the south.

Sporadic sedimentation was continuing in the middle reaches of the streams, although during the so-called Altithermal period streams were cutting channels in the valley deposits. The dating of these intermittent periods of deposition has not yet been worked out to a point where one can say that any particular deposit is so many years of age: all we have thus far is a relative sequence. The exact climate of the Altithermal for this area has as yet not been fully determined; there is conflicting evidence that needs resolution.

A slight rise in effective rainfall occurred about 2000 to 1500 B.C. The people living in this area had so modified their culture that archaeologists refer to material from this later stage as the San Pedro. These somewhat sedentary people constructed houses and storage pits for food stuffs, raised corn, hunted, and gathered. It is worth noting that in several sites along the upper reaches of streams in this general area, San Pedro Stage houses and materials are found resting on a clay that appears to be the same as the "basal" clay of the Blancan Fauna period. No fossils have been found in this particular clay in these localities; hence there is no way of accurately making any correlations between the two clays for age similarities. There are remnants of a soil horizon on this clay horizon which indicates a stability period not present on the other surfaces.

Following the period of circa 1000 B.C. deposition continued in the middle and lower reaches of the streams and gradually covered the habitations of the San Pedro Stage people. Either vegetation in the highlands became sparce and failed to hold the disintegrating rocks, or a change in storm patterns brought on a new set of conditions that released the material to be carried away. This period of intermittent deposition continued until circa 500 B.C.

A long period of erosional stability followed that was probably a result of equilibrium between climate and ground surface, although no well-defined surface has been found thus far that is of this particular age.

At the beginning of the Christian millennia climatic conditions were much

as they are today except that it was probably somewhat drier. This is indicated, for example, by the presence of charred beams of piñon pine (*Pinus edulis*) in archaeological sites whereas today these areas are covered with ponderosa pine (*Pinus ponderosa*). Piñon pine grows at lower elevations and is more resistant to dry conditions than is the ponderosa.

Aside from a short period of deficient effective soil mixture in the early A.D. 300s, conditions remained much the same until about A.D. 550. At this time a decided change in storm patterns caused considerable erosion in the high reaches of the mountain streams, and they began to carry large loads of sediment that they dropped in the middle reaches where there was a break in the stream gradient. In the valleys, canyon bottoms, and large intermontain basins, the depositional material accumulated rapidly. For example, in Chaco Canyon in northwestern New Mexico, recent arroyo cutting has exposed several pithouses more than 15 feet beneath the present surface and others at shallower depths (Judd, 1924; Adams, 1951). In the Rio Grande basin pithouses of this early period are now being exposed that are as much as 10 feet beneath the present surface. Tree-ring dating of these houses gives us an accurate determination of the time periods involved.

During the early 900s, the people in these same areas began constructing large multistoried houses. The surface upon which the houses were constructed is essentially the same as the present day surface, indicating that erosional conditions stabilized about that time, and they have remained essentially stable since then except for the accumulation of wind-blown material and some flood-plain deposits. In certain areas such as some near Gallup, N.M. and in the northern part of the Rio Grande Basin, deposition continued until about A.D. 1250.

During the 1000s and 1100s it was warmer and drier than it had been. Animals from far to the south began to appear in the area, and the people hunted or otherwise killed them, leaving their bones in the midden material belonging to their villages. There was considerable shifting of the human population centers during this period. Large villages in the upper and middle reaches of the basins were abandoned, and the people moved to streams having a more permanent supply of water. Drought has been the favorite reason given for such abandonment.

It was during the period around A.D. 1000 that there was considerable volcanism in the Southwest. The Pinacate field in northwestern Mexico, the field near Verde Hot Springs in Arizona, the San Francisco field near Flagstaff, Ariz., the Grants field near Grants, N. Mex., and the Carrizozo field in east-central New Mexico all contain evidence of lava extrusions and the formation of small cinder cones with accompanying ejectamenta dating roughly from this same time. The small cinder cone near Flagstaff called Sunset Crater is the only one that has been successfully dated, however. Using studies in archaeology, volcanology, ecology, and dendrochronology, I have dated this eruption as having started sometime during the fall, winter, or spring months of A.D. 1064 to 1065 (Smiley, 1958). An interesting part of this eruption is that the ash blown from the crater spread over many square miles of territory and served as a good mulch for "dry-farm" crops in this semiarid area. Shortly

after the eruption many hundreds of people moved into the area to take advantage of the mulching effect of the ash. After about 200 years, winds and other erosional processes piled the ash in dunes, or carried it away until it no longer served its former purpose, and it was found necessary to abandon the area.

During the early 1300s the climate was cooler, and many changes in population centers were made. Some evidence is being discovered that leads me to believe that increasing cold could have brought about this change in the population centers. The cold shortened the growing season of crops so that corn and other staple foods could not be raised, causing widespread starvation. The population dwindled sharply, perhaps due to starvation or disease, or for other reasons as yet unknown. The presence of corn pollen in sediments of this period is much less than it is for older sediments.

The small forested localities in the mountains spread rapidly, and the trees moved out into what had been less favorable areas. Highland types such as ponderosa pine (*Pinus ponderosa*) and Douglas fir (*Pseudotsuga taxifolia*) moved father downward and out into the flat areas, and the piñon-juniper areas spread greatly.

The sharp increase in effective rainfall that began during the 1300s continued until the latter part of the 1500s when it took a sudden and severe drop. Age stands of the forests indicate a widespread increase in the number of trees until the severe period in the late 1500s. This deficient period was primarily a low-altitude drought, at least the evidence is strongest in the low-altitude trees. This effect could have also been caused by a shift from a balanced summer-winter rainfall pattern to more of a summer pattern.

Ideal conditions for tree growth occurred during the 1600s, 1700s, and 1800s, and the forests expanded into areas that had probably not been covered with trees for thousands of years. The smaller pockets of trees that had resulted from the more favorable conditions of the early 1300s served as the seed trees that gave rise to the later forests. Still in existence today are these old age stands wherein the trees reach 600 or more years of age. These pockets are generally in areas of more favorable permanent supply of soil moisture, as on the north slopes of the mountains and in areas of subsurface water drainage.

There has been a sharp reduction in precipitation in the southwest during the last 15 years and, since 1950, trees in many parts of the lower elevations are dying because of starvation and parasitic activity. The forests are pulling back because of this decrease in moisture. Whether this is a definite trend or a short range fluctuation is yet to be determined.

References

Adams, R. N. 1951. Half House: a pithouse in Chaco Canyon, New Mexico. Papers of the Michigan Academy of Science, Arts, and Letters. **35:** 273–306.

Antevs, Ernst. 1959. Geological Age of the Lehner Mammoth Site. American Antiquity. **25**(1): 31–34.

Broecker, W. S. & P. C. Orr. 1958. Radiocarbon chronology of Lake Lahontan and Lake Bonneville. Geol. Soc. Am. Bull. **68:** 1009–1032.

Clisby, K. & P. B. Sears. 1956. San Augustin Plains—Pleistocene climatic changes. Science. **124**(3221): 537–539.

Damon, P. E., B. Sables & A. Long. 1960. Geochemistry. *In* The Utilization of Arid

Lands, An Interdisciplinary Study. : 17–34. 2nd Ann. Rept. Univ. Ariz. Tucson, Ariz.

de Martonne, E. 1948. Traité de Géographie Physique. 7th ed. Librairie Armand Colin. Paris, France.

Haury, Emil W., E. B. Sayles & William W. Wasley. 1959. The Lehner Mammoth Site, Southeastern Arizona. American Antiquity. **25**(1): 2–30.

Hettner, A. 1930. Die Klimate der Erde. Teubner. Leipzig, Germany.

Judd, N. M. 1924. Two Chaco Canyon pithouses. Ann. Rept. Smithsonian Institution for 1922. : 399–413.

Köppen, W. 1936. Das geographische system der Klimate. **1:** Part C. Berlin, Germany.

Lance, John F. 1959. Faunal Remains from the Lehner Mammoth Site. American Antiquity. **25**(1): 35–42.

Martin, P. S., B. Sables & D. Shutler, Jr. 1961. Pleistocene ecology and extinction of the Shasta Ground Sloth. Am. J. Sci.

Smiley, T. L. 1958. The geology and dating of Sunset Crater, Flagstaff, Arizona. *In* Guidebook of the Black Mesa Basin, Northeastern Arizona. R. Y. Anderson and J. W. Harshbarger, Eds. N. Mex. Geol. Soc. : 186–190.

Smiley, T. L. 1959. Paleoecology of the Cochise culture area, southeastern Arizona. To be published as part of a memoir in American Antiquity.

Thornthwaite, C. W. 1933. The climates of the earth. Geogr. Rev. **23:** 433–440.

Thornthwaite, C. W. 1948. An approach toward a rational classification of climate. Geogr. Rev. **38:** 55–94.

Von Eschen, G. F. 1958. Climatic trends in New Mexico. Weatherwise. **11**(6): 191–195.

CLIMATIC CHANGES AND PREHISTORIC AGRICULTURE IN THE SOUTHWESTERN UNITED STATES*

Richard B. Woodbury
University of Arizona, Tucson, Ariz.

Although the term Southwest has never been defined to everyone's satisfaction, it continues to be a convenient way of referring to a part of North America that has a relatively homogeneous and distinctive aboriginal cultural development. Used in this cultural-geographical sense it refers to the Great Basin, the Southern Rocky Mountains, the deserts of Arizona, New Mexico, Sonora, and Chihuahua, and the northern part of the Sierra Madre. In spite of the great variations in topography, climate, and vegetation found within this region, it has been the scene of only two main streams of aboriginal cultural development in the past four or five millenniums. The first was based on an economy of gathering wild plants and hunting, with emphasis on small game; the second was based on agriculture. The first of these types of subsistence was a gradual development, rooted in the old-world background of the first migrants to the New World. Prior to their extinction, large game animals were a major food resource of at least part of the population, but small game and wild plants were also important even to the mammoth hunters. As a result of the disappearance of the large mammals the less spectacular food resources became of primary importance. This economy, based on wild plants and small animals, persisted of course into historic times for many groups within the southwestern region.

In spite of the wealth of detail that has been accumulated in nearly a century of archeological excavation, a satisfactory up-to-date account of the major trends and important regional differentiations is not available. A summary compiled in 1955 (Jennings, 1956) can be recommended, however, for filling in the picture sketched here in greatly simplified form.

The widespread and ancient collecting economy of the Southwest (sometimes called the "desert culture" and also identified as part of the continent-wide "archaic") gradually gave way to a farming economy, but only in the more suitable areas and in them only partially. In addition to maize, both squash (*Cucurbita*) of several species and beans (*Phaseolus*) were introduced from Mexico, and the three plants became the staple diet of farming groups, supplemented by many wild products. The introduction of pottery and the development of several types of fairly permanent houses, built in increasingly large village clusters (Willey, 1956), were accompanied by marked regional differentiation. A culture climax was reached in the centuries between 700 and 1200 A.D., with agriculture practiced over its greatest extent, population at its peak, and the material manifestations of cultural vigor achieving notable local diversity. Shortly before the end of the 13th century a rapid withdrawal began that left larger and larger parts of the Southwest unoccupied by farming peoples (Reed, 1944). In so far as the archeological evidence goes, these areas were not permanently inhabited again until shortly before historic times when

* The research on which this paper is based was made possible by the Arid Lands Research Program (Contribution No. 39) of the University of Arizona, supported in part by The Rockefeller Foundation, New York, N.Y.

Athabascans arriving from the north occupied some of them. The San Juan area was among the first abandoned, along with much of the lower part of the Little Colorado River area and a large area of northwestern New Mexico. In the late 14th century and during the 15th century, depopulation extended to the mountainous region of east-central Arizona and west-central New Mexico, the upper Little Colorado except for a small area around Zuni, a portion of the Rio Grande drainage in southern New Mexico, and the Salt and the upper Gila River drainages. Thus in less than 3 centuries the area occupied by farmers who had made an apparently successful exploitation of a variety of environments shrank from an estimated 230,000 sq. mi. to about 85,000; most of this smaller area was south of the present Mexican border, where the aboriginal occupation had never been as dense as farther north. Although the last 2 centuries before the Spanish conquest saw a continuation of the same farming cultures and even an elaboration of certain architectural, religious, and artistic traits, there was a further shrinkage of the area occupied that has continued to the present century.

The introduction of agriculture into the Southwest made possible the development of the village form of life; the population density and the specialized, complex activities set this area apart from neighboring regions to the west, north, and east. Moreover, since agriculture depends intimately on several environmental factors, climate among them, the evidence for its appearance, gradual growth in importance, and rapid decline deserves attention. Farming seems to have come into the area by means of a gradual northward spread through the northern Sierra Madre into the higher, more mountainous portions of the Southwest. The oldest evidence is from Bat Cave, overlooking the San Agustin plains of western New Mexico, where tiny ears of a primitive type of corn have been dated by the C-14 method to about 3500 B.C. (Mangelsdorf, 1958), when the end of the Altithermal brought cooler and moister conditions to the area. Considerably later there is evidence of corn (identified only by means of pollen) at Point of Pines in eastern Arizona (Martin, 1960). These locations are about 125 miles apart, between 6000 and 7000 feet above sea level, in the mountainous zone of the Southwest that lies between the southern deserts and the northern plateaus. During the Altithermal many areas of lower elevation may have been scantily populated or even unoccupied, whereas the mountain zone would have continued to offer a variety of plants and animals. The small size of the ears of this early corn and the probable poor yield from the plants make it improbable that it was at first more than a minor addition to the hunting and collecting economy. This inference is supported by the careful analysis of the proportion of the food supply that is represented by different kinds of remains in Tamaulipas, over 1000 miles to the southeast in Mexico (MacNeish, 1958): in the La Perra phase, at about 300 B.C., about 20 per cent of the food energy of the group is estimated to have come from agriculture, the rest about equally from wild plants and from wild animals. In the Bat Cave and Point of Pines areas, farming probably advanced even more slowly than this. It may be of some significance that these agricultural beginnings in New Mexico and Arizona come at the close of the Altithermal since, during the cooler climatic period that followed for about 2000 years (except for one warm oscillation), we have no further evidence of agriculture. This may

be due to the paucity of archeological evidence from this early period when sites are few, inconspicuous and, for the most part, destroyed by subsequent natural events. However, it may also be that agriculture failed to spread or even to survive, and was reintroduced from the south a few centuries before the beginning of the Christian era, when somewhat warmer conditions again prevailed. In Tularosa Cave, in deposits dated at about 300 B.C., large quantities of maize have been found, suggesting an increasing dependence on farming. This site is located about halfway between Bat Cave and Point of Pines, so that it is possible that corn persisted in this limited region and here finally began to assume its dominant role in the economy. From this time onward corn was of increasing importance and was grown in a rapidly widening region, northward across the Colorado plateau into southern Colorado and far up into Utah, southwestward into the valleys of the Gila and Salt rivers and of their tributaries (but never very successfully in the deserts flanking the rivers), and eastward in New Mexico, reaching the Rio Grande, the Pecos and, finally, the Upper Canadian rivers.

The changing climatic situation that accompanied this agricultural development can be briefly summarized (partly on the basis of a paper by R. W. Fairbridge presented in New Haven, Conn., on May 7, 1960, at the 25th Annual Meeting of the Society for American Archaeology). The Altithermal was followed, as has been mentioned, by a long period of cooler conditions, interrupted by a somewhat warmer period of roughly 1700 to 1500 B.C. By 300 B.C., warmer conditions had again returned. A cooling trend occupied the next 3 or 4 centuries, but from about 100 to 1100 A.D. there was increasing warmth, apparently on a worldwide scale (reflected, for example, in the Rottnest Terrace; Fairbridge, 1958). The 12th century was marked by a rather rapid change to cooler conditions, and the 13th century by a major drought, widely reflected in tree rings in western North America (Schulman, 1956). About 1600, a minor warm oscillation brought conditions temporarily to approximately the same temperatures that have again been reached in the present century.

Since agriculture reached its maximum about 1100 A.D., a time of relatively warmer and probably drier conditions, the long-held idea that drier periods in the Southwest were unfavorable to farming would seem untenable. A compensation for moderate decreases in precipitation may have been achieved by means of the increasingly extensive and elaborate techniques that were practiced for the control of erosion and the conservation of surface runoff on gently sloping surfaces, by the systems of terraces on steep, intermittent streams, and by ditches for irrigation along stream bottoms. These water-control structures occur more widely in the Southwest than has been generally recognized, and their greatest development began in the 2 centuries following 900 A.D. (Woodbury, 1961), a time of declining moisture. By making more efficient use of the available moisture and by making farming practical in many locations hitherto unused, these new farming techniques may have gone a long way toward offsetting the adverse effects of climatic deterioration. Therefore any interrelation between climate and agriculture must be viewed in the light of the technological possibilities that would provide the means of adapting farming practices to changed environmental conditions.

Another little-noticed possibility in considering the delicate balance that has existed between man and nature in the arid western portion of North America is the significance of temperature, as manifested in the length of the growing season. A moderate warming of climate, such as seems to have occurred with the maximum spread of agriculture, would make farming feasible at somewhat higher elevations. In fact such an expansion would have been impossible until late in the first millennium B.C. and, although begun then, it may have been retarded for several centuries by the cooler climate of the first half of the first millennium A.D. An example will make clearer the close relation between elevation and growing season. An examination of recent weather records at Zuni, N. M., and at El Morro National Monument about 30 miles to the east and nearly 1000 feet higher, show that the average annual temperature at El Morro is from 2° to 6° F. lower, and that the growing season (days between the latest spring and earliest autumn temperatures of 32° F.) is from 3 to 78 days (average 30 days) shorter. In terms of aboriginal agriculture this longer growing season at Zuni would have been of crucial importance. The extension of the growing season at higher elevations (where moisture is often a little greater) as a result of widespread warming was probably an important stimulus to the general expansion of agriculture until its maximum extent was reached in the 12th century.

If the relationships just mentioned, between elevation and growing season, are valid, the abandonment of large parts of the Southwest by farmers that began in the 13th century may be partially explicable in terms of them. Archeologists are now agreed that theories of disease and of invasion by nomads are entirely inadequate to explain such a widespread decline in population. Drought has been widely cited as the major causative factor (Kelley, 1952; Schwartz, 1957), and with considerably more justification, since the dendrochronological and the physiographic records furnish convincing evidence of repeated droughts, and rain-making ceremonies still survive conspicuously among the descendants of the Southwest's aboriginal population. Besides its direct effect on growing crops, drought had a significant effect in some regions through its initiation of arroyo cutting and the consequent lowering of water tables and speeding of runoff (Bryan, 1941; Reed, 1944). However, the lowering of temperatures that accompanied the Southwest's agricultural contraction may have had an equally drastic effect, and one that would be felt promptly in areas that were not affected until later, if at all, by arroyo cutting. Also a shortening of the growing season, with its devastating effect on agricultural productivity, could not be even partly compensated for by any such simple but extensively used means as terraces or stone alignments, and the increased rainfall that might accompany the cooling trends would be useless if frost prevented crops from maturing. The aboriginal practice of planting fields in a variety of locations, with different exposures, water supply, and drainage, also served as a partial insurance against losses from extreme precipitation fluctuations, but was less helpful in reducing frost damage.

It should be noted that in general the more northerly and higher areas tended to be abandoned first, even though they were favored by somewhat greater rainfall than many of the locations where occupation continued. However, such observations rest on quite inadequate data, and the preceding discussion

should be viewed as programmatic rather than conclusive. What is needed is a careful study of the interrelations of cultural and environmental factors in the growth and decline of aboriginal farming in the Southwest, with consideration of such factors as elevation and growing season, and of water control by aboriginal engineering techniques. A valuable contribution could be made by a careful mapping of the progressive abandonment, century by century, of farming areas, in relation to elevation as well as to arroyo cutting. A perfect correlation between withdrawal and elevation, or any other environmental factor, is impossible, since groups varied in such cultural attributes as ability to conserve food, and in reluctance to abandon their homelands, and areas varied in the nonagricultural resources they offered. Along with such mapping we greatly need additional study of the climatic events of the past two millenniums in the Southwest in order to define the extent to which they varied from world-wide trends. This will provide a more satisfactory basis for estimates of the temperature and precipitation changes that would have affected the aboriginal farming population. With the extensive archeological information that has been accumulated for the area, and an extremely close chronological control for the data, a particularly promising opportunity is provided for reaching a better understanding of the intimate and complex interplay of natural and cultural forces in shaping the life of man.

References

Bryan, K. 1941. Pre-Columbian agriculture in the Southwest, as conditioned by periods of alluviation. Ann. Assoc. Am. Geog. **31:** 219–242.

Fairbridge, R. W. 1958. Dating the latest movements of the Quaternary sea level. Trans. N. Y. Acad. Sci. Ser. II. **20**(6): 471–482.

Jennings, J. D. (Ed.). 1956. The American Southwest: A problem in cultural isolation. Soc. Am. Archaeol. Mem. **11:** 59–127.

Kelley, J. C. 1952. Factors involved in the abandonment of certain peripheral southwestern settlements. Am. Anthropol. **54:** 356–387.

MacNeish, R. S. 1958. Preliminary archaeological investigations in the Sierra de Tamaulipas, Mexico. Trans. Am. Phil. Soc. (n.s.). **48**(6): 1–210.

Mangelsdorf, P. C. 1958. Ancestor of corn. Science. **128:** 1313–1320.

Martin, P. S. 1960. Arizona's oldest cornfield. Science. **132:** 33–34.

Reed, E. K. 1944. The abandonment of the San Juan region. El Palacio. **51:** 61–74.

Schulman, E. 1956. Dendroclimatic changes in semiarid America. Univ. Ariz. Press. Tucson, Ariz.

Schwartz, D. W. 1957. Climate change and culture history in the Grand Canyon region. Am. Antiq. **22:** 372–377.

Willey, G. R. (Ed.). 1956. Prehistoric settlement patterns in the New World. Viking Fund Publs. Anthropol. **23.**

Woodbury, R. B. 1961. Prehistoric agriculture at Point of Pines, Arizona. Soc. Am. Archaeol. Mem. **17.**

SOME CORRELATIONS OF CLIMATIC AND CULTURAL CHANGE IN EASTERN NORTH AMERICAN PREHISTORY

J. B. Griffin
University of Michigan, Ann Arbor, Mich.

There is thus far relatively little direct evidence for climatic change from archaeological sites in the eastern United States. Archaeologists have considered for many years that the dividing line between eastern and western is the eastern side of the Rocky Mountains. This whole area then, has become a convenient unit of prehistoric studies because of the obvious connections and similarities at all known time periods and all of the cultural levels. Since A. D. Krieger, in his paper in this volume, emphasizes the initial occupation in North America, my attention will be primarily devoted to the last 4000 to 5000 years.

Before one can interpret cultural changes as resulting from climatic variation it is highly desirable to have one's prehistoric house in reasonable order. Sound chronological and stratigraphic sequences need to be developed over a wide area, and cultural complexes need to be carefully defined in local areas and in their correct and reasonably precise temporal position in order to observe the effects of climatic change on cultural behavior. Not only is a sound and laboriously obtained archaeological framework necessary, but there needs to be a sound corpus of data regarding floral and faunal changes, of soil zones, and of other climatic indicators in general ecological areas or zones. The pollen studies of the past 30 years have been a great help, and radiocarbon dating is providing an increasingly accurate chronological scale.

Three years ago, while working on a paper on the prehistory of the Winnebago tribe of Wisconsin (Griffin, 1960*b*), I developed the thesis that the observed cultural decline in the Upper Mississippi Valley of the Winnebago and their ancestors could be explained, at least to some degree, as the result of a climatic deterioration that corresponded to that accurately documented in Northern Europe and Greenland from about A.D. 1200 to 1700. Beginning a few hundred years before A.D. 1000 there developed in the East St. Louis area an agricultural, sedentary culture type that is a regional expression of an increasing emphasis on agriculture known in much of the eastern area as the Mississippi culture (Griffin, 1952). The Old Village complex of Mississippi culture in the St. Louis area exerted a strong influence to the north and northwest, up the Illinois, Mississippi, and the Missouri valleys. It is certain that there was a strong movement of people into southern and western Wisconsin by about A.D. 700 to 800 (M-1037) and into eastern Minnesota, taking with them their customary cultural characteristics and their predominantly agricultural economy. Somewhat similar but apparently less strong influences moved into northwest Iowa and southeastern South Dakota.

From the archaeological evidence, we had known for some time that the late prehistoric and early historic culture of the Upper Mississippi area, called Oneota, was a descendant of the Old Village tradition (Griffin, 1943, p. 302) that had shifted away from a marked dependence on agriculture to a heavier

emphasis on hunting, and that there was a marked drop in cultural level. This period of gradual decline can now be clearly seen as occurring in the A.D. 1300 to 1650 time period. Preliminary reports of the Mill Creek culture in northwest Iowa also indicate a gradual shift toward increased emphasis on hunting from the Old Village influenced lower levels of deep middens, to the upper and later levels that should date from A.D. 1400 to 1600.

In the lower Great Lakes and New York area an increasing dependence on agriculture and a more sedentary way of life begins during the A.D. 700 to 900 period. This regional development does not go into a decline but instead seems to reach a climax in the Iroquoian culture type in lower Ontario and New York between A.D. 1400 to 1600. It is suggested that the modifying effect of the Great Lakes on this area was sufficient to prevent a decline in effectiveness of agricultural crops and thus sustained the Iroquoian economy. Such an effect is in operation now on the present flora (Soper, 1956).

In considering the archaeological evidence from the plains it is necessary to keep in mind the marked division in that area between the eastern and western plains. This division is recognized in vegetation and animal life, which reflect climatic differences, primarily rainfall, which decreases markedly from Omaha to Denver. By A.D. 700 to 900 ceramic producing and partially agricultural people have moved, or their culture pattern has spread, from the eastern plains of Kansas and Nebraska into eastern Colorado and the Texas and Oklahoma panhandles. Here they come into contact with peoples moving eastward from the Southwest culture center. From A.D. 900 to 1300 there are settled agricultural villages in the high plains located on well developed humus zones that overlie soils of a drier period (Wedel, 1941, p. 19). Between A.D. 1300 to 1500 there is a marked withdrawal of these people to central and eastern South Dakota, Nebraska, Kansas, and Oklahoma (Wedel, 1953, figure 2). In this area they are strongly concentrated in their agricultural villages along the major stream valleys with sources in the Rocky Mountains. The sites in the high plains which have been abandoned, are covered by wind blown soils. This is not the case with the abandoned sites of the same period in the eastern plains.

Turning briefly to the Southwest we see, on maps prepared by Stanley Stubbs, that the period from A.D. 700 to 1200 is marked by a great expansion of agricultural village settlements from an initial center in the four corners area, to occupy much of Utah as far as the area of Great Salt Lake, into eastern Colorado and eastern New Mexico (Jennings, 1955, pp. 64 to 66). Shortly before A.D. 1200, a marked retraction sets in, so that by the time of the Spanish penetration into the Southwest the Pueblo agricultural Indians were concentrated along the Rio Grande and in other local areas where there was arable land and a dependable water supply from high-altitude sources. A similar expansion and retraction took place in southern Arizona where a striking prehistoric culture, the Hohokam, flourished and developed an extensive irrigation system. It should also be noted that in northwestern Mexico, along the eastern slope of the Sierra Madre range, there was a notable growth of agricultural settlements in the 700 to 1350 period, and that during the following centuries there was a marked abandonment of these sedentary sites (Kelley, 1956, p. 132). There are a

number of significant evidences of erosion in the Southwest from 1200 to 1700 in the alluvial deposits built up during the general period of alluviation of the A.D. 700 to A.D. 1100 period.

It is believed that the period from A.D. 700 to 1200 in the Southwest and western plains is a period of more adequate and equitable distribution of rainfall that would allow the growth and expansion of argicultural communities. The temperature conditions might or might not be significantly different from those of the present. In this same period there seems to be a warmer climate in the Upper Mississippi Valley and Great Lakes which allows the settled agricultural populations to move northward into locations near the present border of effective corn agriculture.

The 1200 to 1700 period in the Great Lakes and northeast is believed to be relatively cooler and with a shorter growing season. It is probably also somewhat moister. It is my belief that minor changes in pollen diagrams from Minnesota to Maine reflect this relatively colder period. Minor glacial advances are recorded for this period, and the recent sea-level rise is from the lower levels reached because of glacial growth. As observed above, it is a period of increasing cold and storminess in Northwestern Europe that is historically documented. In the Southwest and western plains it is evident that this period is one of increased dryness and of marked erosion.

By dropping backward another 1000 years we can see a similar sequence that parallels that from A.D. 700 to 1700. Between approximately 300 B.C. to A.D. 300 there developed in the Ohio and Illinois areas a culture called Hopewellian that is sedentary, ceramic-using and, to some degree, agricultural (Griffin, 1958). From the central Illinois area, Hopewell people and cultural influences move into northern Indiana and Michigan, to as far north as the Muskegon-Saginaw line. There is also a movement up the Mississippi into Wisconsin and eastern Minnesota, as far as the mouth of the Minnesota and to Green Bay. Cultural spread also takes place to eastern Kansas, Nebraska, and northeastern Oklahoma. The westward movement of people on a Hopewellian level reaches western Kansas and perhaps to eastern Colorado.

The Hopewellian development was a remarkable one in many ways, and it it was their cultural products, earthworks, and elaborate burial mounds that was largly responsible for the term Mound Builders. That term, incidentally, was abandoned by archaeologists many years ago. For many years archaeologists have been puzzled by the disappearance of the Hopewellian culture. Recently we have been able to recognize that it did not suddenly disappear by conquest or migration but that, instead, it gradually declined into a simpler and less spectacular cultural form. In every area where it was dominant, from eastern Ohio to Kansas City and from Illinois into Wisconsin and Michigan, there is a gradual change into Late Woodland Culture expressions which can now be accurately dated from A.D. 300 to 400 to 700 to 1000. In the areas north and east of the Mississippi cultures this general Late Woodland Culture was maintained to the period of European exploration and colonization.

Last spring I prepared a short paper postulating that the Hopewellian growth and spread to the north coincided with the evidence from northwestern Europe of a relatively warm-dry period, and with evidence for a sea-level rise representing a general climatic moderation (Griffin, 1960*a*). This same period of 300

B.C. to A.D. 300 marks the initial development in the Southwest of small, settled village agriculturists in southwestern New Mexico and southeastern Arizona, and of the period of growth of the Basketmaker II culture in the four-corners area of the Southwest. At present I am not aware of much evidence of a climatic deterioration in the Southwest in the next 400 to 500 years, but Antevs (1955, p. 330) has called attention to the Whitewater drought and erosion in the Southwest, and this is also reflected in the tree-ring pattern. At some sites in the western plains, of the Hopewellian level, which is on a well-developed humus zone, there is a soil cover over the village site indicating a period of markedly drier climate before the later humus soil of the A.D. 800 to 1200 period (Wedel, 1941).

The alternating climatic pattern described for the period of A.D. 700 to 1700 is thus foreshadowed by that from about 300 B.C. to A.D. 700 in the Southwest, in the western plains, and the Upper Mississippi Valley-Great Lakes. A relatively mild climate in the latter area coincides with effective rainfall for vegetation and agriculture in the Southwest and western plains. A colder climatic regime in the north and northeast coincides with evidences of drought and erosion in the Southwest and evidence of wind blown soils in the western plains.

There is very little direct evidence on climate from archaeological sites in the period from 2000 B.C. to 300 B.C. in the eastern United States. If the pattern of alternating cool and warm periods continues in its suggested regularity into the past, then the period from ca. 800 to 300 B.C. should be relatively cool; from 1300 to 800 B.C., relatively warm; 1800 to 1300 B.C., relatively cold and, from 2300 to 1800 B.C., relatively warm. We should also find, in the Southwest and western plains, indications of alternating drought and alluvial deposition. The latter accompanied by a more favorable rainfall pattern for vegetation.

The evidence from the Southwest is not clear, uniform, and accurately dated in the majority of cases. There are also gaps in the areas studied and either in the formations themselves or in the interpretive record. It is also true that if erosion is operative in a particular local area that such eroded material will be deposited in another area. Antevs (1955, figure 93) has called the Fairbank Drought a period of drought and erosion dated at 500 B.C. There are other loci in the Southwest and the high plains that indicate that this general period is relatively dry. The Hopewellian period humus zone occupation in western Kansas of about A.D. 100 to 200 is underlain by a sterile grey windblown (?) soil that may well belong to the 800 to 300 B.C. period of drought.

The Fairbank Drought is preceded by a period of alluvial and colluvial deposition that is placed on a recent correlation about 200 years on each side of 1000 B.C. (Haynes and Agogino, p. 21). I know of no supporting evidence from the Northeast to indicate that the period of 1300 to 800 B.C. actually has a warmer climate. There is, however, a warm dry spell clearly recorded in England at this time (Godwin and Willis, 1960, p. 66, Q-389), and the same general conditions are expectable in the Great Lakes. There is a raised beach line 4 feet above sea level in New Zealand that is documented by Schofield (1960, p. 479), and this should be the correct period for the Younger Peron Terrace of Fairbridge (1958). This is a cultural period in the Great Lakes and

Northeast in which we recognize a striking development of burial ceremonialism spread from southern New England to the northern Wisconsin area. Some of the cultural characteristics and, perhaps, people as well may have moved south into Illinois and the Ohio Valley during the cold period of 800 to 300 B.C. and, if so, they helped to stimulate the striking cultural development in that area of the next warm period: one that, however, is clearly forming to the south during the 800 to 300 B.C. level. This cool-moist period in England is represented by the trackways of the Late Bronze Age in Somerset dated between 870 and 310 B.C. that were laid to allow crossing of flooded bogs (Godwin and Willis, 1959 and 1960). The earlier date is also that of the *Grenzhorizont.*

An excellent example of a minor effect on cultural patterns in the Archaic stage in the Southeast may be seen in the headwaters of the St. Johns River, and on Indian River in east-central Florida, where sea-level changes caused minor population displacement but no significant cultural change. In this area the Late Archaic preceramic complex and the early ceramic period is found on the relatively dry land surface of the Melbourne-Van Valkenburg interval. Rouse (1951, pp. 239–263) has an excellent summary in which the relations of the sites to changes in sea level and environment are emphasized. He believed, and still believes (personal communication, January 1961) that the date of the early ceramic level corresponds to approximately 1000 B.C. I am not certain, however, as to the precise correlation of Rouse's cultural and climatic interpretations with the fluctuations of sea level of Fairbridge. The culmination of the Younger Peron Terrace is given about 1700 B.C. which is probably too early for the Florida data of Rouse. Certainly, radiocarbon dating is desirable for the specific cultural complexes involved. The sea level changes that have been documented thus far will certainly be revised and probably increased as new and more accurate assessments and interpretations are made. I believe that something is wrong with the spread of years given by Fairbridge for the Younger Peron Terrace. The sea-level rise of Rouse's early ceramic period is probably one that occurs about 1200 B.C.

I shall skip over the postulated cold period in the north from 1800 to 1300 B.C. for which there is some documentation in the form of lowered sea levels, peat deposits, and trackways in England, and a presumed, on my part, drought and erosion in the Southwest. For the postulated warm period in the north from ca. 2300 to 1800 B.C. there is a considerable amount of confirmatory evidence. The post-Glacial rise to the period of maximum warmth seems to reach its peak in this period with the most northern extent of the deciduous forest. The general post-Glacial northward movement of the flora with its accompanying animal life allowed the northward movement of prehistoric culture complexes that had developed in this ecological zone, in the Ohio Valley, in the Central Mississippi Valley, and along the Middle Atlantic Coast. It is quite clear that this northward movement of culture did take place and that from Wisconsin, across the lower Peninsula of Michigan, across Ontario, and into New York and New England, there is a marked flowering of what we call Archaic cultures in this Late Archaic period. This development marks the beginning of the Old Copper Culture in the western Great Lakes and of the development of the Lamoka, Laurentian, and coastal cultures of the Northeast.

These cultures did not originate in or develop in a "boreal" forest zone, and only rarely did they move into one.

For this general cultural level we have recorded at some sites the presence of acorns and hickory nuts. We know that a deciduous forest covers the area delineated above by this time. The importance of these nuts, as food, is emphasized by the numbers of grinding stones used for their preparation in the Lamoka culture (Ritchie, 1932). The most important food supply, however, appears to have been derived from hunting and fishing. During this period of time, 2500 to 2000 B.C., we find a rather stable occupation by Indian groups at favorable sites over a fairly long period. A specific indication of Lamoka period and climate is the presence of the fresh-water snail (*Galba catascopium*) in the bottom level of the Lotus Point Site, N. Y., with a complex of the Lamoka culture (Ritchie, 1958, p. 33). The habitat of this snail is such that it is suggested that the Hudson River level was 10 to 12 feet higher than at present. The lower levels of the river site, with an "early" Lamoka occupation, were interpreted geologically as belonging to a period of higher water level in the Hudson (ibid., p. 44). This would seem to be best interpreted as that of the high water stage of about 2000 to 1800 B.C. represented by raised beaches in Algeria and England, or, less likely, the Older Peron Terrace of 3100 B.C. (Fairbridge, 1958).

The date for the sub-Boston Boylston Street fish weir and its burial by marine silts is presented in this paper as representing the period of a major rise in sea level, about 2000 B.C. Underlying the fish weir, which was built in the mud flat and marsh land, was a 9-inch layer of fresh-water peat, which has a possible shift from a drier phase in the lower section and a moister phase toward the top (Benninghoff, 1942, p. 104). A sample of this peat from an unspecified section of the deposit is dated at 3762 B.C. $\pm$ 500 (C-417). One of the preserved stakes, of an estimated 65,000, has a date of 4500 $\pm$ 130 or ca. 2500 B.C. About 4 feet above the fish weir, in marine silts, was a stream rafted stump that has been given a date of 1896 B.C. $\pm$ 390 (C-418). The exact meaning of these dates, in terms of the complex relationship between sea and land levels in the Boston area, is left to those who have actively participated in its interpretation. It is inferred here that the sea was rising and that its rise reflects that observed in other areas. It is clear that the vegetational picture of the lower peat and of the time of the weir indicates a climate comparable to, and then warmer than, that of the present time (Barghoorn, 1959, pp. 109 to 110).

The cultural pattern of the fish-weir builders would be very close to that of the Late Archaic occupants of the Wapanucket site near Middleboro, Mass. (Robbins, 1959), where three radiocarbon dates (M-764, 969, and W-363) place the period at 2300 B.C. From other evidence in southern New England we know that this period marks a climax of the Archaic occupations in terms of stability and number of sites. This period of 3000 to 2000 B.C., of northward movement of developed Archaic complexes, is also the one in which there is good reason on archaeological grounds to believe that the Minnesota Man or Lady belongs.

Summary

The above discussion has focused attention on some proposed climatic fluctuations in the southwestern United States, the upper Mississippi Valley-Great Lakes, northeastern United States and northwestern Europe. These alternations of climatic conditions appear to occur between about 500 to 600 years. The period chosen for discussion of the last 4000 years is one for which we have some adequate documentation and for some of which we have suggestive historical records. The presentation made in this paper should be viewed as schematic and provocative. It will need more adequate documentation and temporal control to be adopted as a working hypothesis for intensive study.

The approximately 550 years periodicity of climatic change proposed above is one recognized by other students and for which there is evidence in other areas and at other times. In Denmark and North Germany, Barendsen *et al.*, (1957, Figure 1) have recently presented tight radiocarbon dating for minor climatic fluctuations of ca. 550 years, representing changes from continental to maritime climates during the period from ca. 11,660 B.C. to 8500 B.C. These alternations represent climatic fluctuations of a somewhat similar pattern in the minor advances and retreats of the glaciers during their gradual withdrawal from the upper Great Lakes and lower Canada.

The implication of the periodic changes in climatic patterns for the three areas discussed in this paper is that the changes were probably brought about by changes in the circulation pattern of the earth's atmosphere. I have discussed this question with certain meterologists who will remain nameless, and thus blameless, for any misunderstanding of their views. One suggestion has been that cool-moist conditions in the northern and northeastern United States and evidence of drought in the Southwest is probably produced by a low-latitude zonal circulation that also causes climatic stress, such as the development of heavy storms along the eastern part of the country and in the North Atlantic and Europe. On the other hand a high-latitude zonal circulation would produce warm and relatively dry conditions in the north and northeast and would bring greater summer precipitation in the Southwest and the western plains.

If the zonal circulations can be shifted from north to south in their general movement, so that for the period suggested, ca. 550 years, they have primarily high or low zonal movements, what is the controlling mechanism, and how is the suggested periodicity explained? If additional and more careful analysis and synthesis of the available paleoclimatic data tend to substantiate the pattern proposed here, then, presumably, the total air circulation patterns of the world must have produced this effect. Similar climatic shifts should be discernible in other areas where temperature and drought have been controlling factors in vegetational growth, with resultant affect upon human culture. One should be able to predict, for example, the periods at which north and northwestern China were favorable for agricultural production and which were not, and the periods during which the Mediterranean basin and North Africa were most favorable for agriculture and extensive human occupancy. It would be rather interesting to know.

References

ANTEVS, E. 1955. Geologic-climatic dating in the West. Am. Antiquity. **20**(4). Pt. 1. : 317–335.

BARENDSEN, G. W., E. S. DEEVEY & L. J. GRALENSKI. 1957. Yale radiocarbon measurements III. Science. **126:** 908–979.

BARGHOORN, E. S. 1955. Paleobotanical studies in salt marsh deposits with special reference to recent changes in sea level. *In* Proceedings, Salt Marsh Conference. : 109–110. The Marine Institute. Univ. Georgia. Athens, Ga.

BENNINGHOFF, W. S. 1942. The pollen analysis of the Lower Peat, in the Boylston Street fishweir. Papers of the R. S. Peabody Foundation for Archaeology. **2:** 96-104.

FAIRBRIDGE, R. W. 1958. Dating the latest movements of the Quaternary sea level. Trans. N. Y. Acad. Sci. Ser. II. **20**(6): 47–482.

GODWIN, H. & E. H. WILLIS. 1959. Cambridge University natural radiocarbon measurements I. Am. J. Sci. Radiocarbon Suppl. **1:** 63–75.

GODWIN, H. & E. H. WILLIS. 1960. Cambridge University natural radiocarbon measurements II. Am. J. Sci. Radiocarbon Suppl. **II:** 62–72.

GRIFFIN, J. B. 1943. The Fort Ancient Aspect: Its cultural and chronological position in Mississippi Valley archaeology. Univ. Mich. Press. Ann Arbor, Mich.

GRIFFIN, J. B., (Ed.) 1952. Archaeology of Eastern United States. Chicago Univ. Press. Chicago, Ill.

GRIFFIN, J. B. 1958. The chronological position of the Hopewellian culture in the Eastern United States. Anthropological Papers, No. 12, Museum of Anthropology. Univer. Mich. Ann Arbor, Mich.

GRIFFIN, J. B. 1960*a*. Climatic change: a contributory cause of the growth and decline of northern Hopewellian culture. Wis. Archaeologist. **41**(1): 21–33.

GRIFFIN, J. B. 1960*b*. A hypothesis for the prehistory of the Winnebago. *In* Culture in History, S. Diamond, Ed. : 809–868. Columbia Univ. Press. New York, N.Y.

HAYNES, V. & G. AGOGINO. 1960. Geological significance of a new radiocarbon date from the Lindenmeier Site. Proceedings No. 9, Denver Museum of Natural History. Denver, Colo.

JENNINGS, J. D., (Ed.) 1955. The American Southwest: A problem in cultural isolation. *In* Seminars in Archaeology. : 59–127. Memoir II, Society for American Archaeology. Salt Lake City, Utah.

JOHNSON, F. S. 1942. The Boylston Street Fishweir. Papers of the R. S. Peabody Foundation for Archaeology, Vol. 2, Andover.

JOHNSON, F. S., (Ed.) 1949. The Boylston Street Fishweir II. Papers of the R. S. Peabody Foundation for Archaeology. **4**(1), Andover.

KELLEY, J. C. 1956. Settlement Patterns in North-central Mexico. *In* Prehistoric Settlement Patterns in the New World, G. R. Willey, Ed. Viking Fund Publications in Anthropology. No. **23:** 128–139. New York, N.Y.

RITCHIE, W. A. 1958. An Introduction to Hudson Valley Prehistory. New York State Museum and Science Service Bull. No. **367**. Albany, N.Y.

RITCHIE, W. A. 1932. The Lamoka Lake site: The type station of the Archaic Algonkin period in New York. Researches and Trans. N. Y. State Archeological Assoc. **7**(4): 79–134.

ROBBINS, M. 1959. Wapanucket No. 6: An archaic Village in Middleboro, Massachusetts. Cohannet Chapter, Mass. Archaeological Soc.

ROUSE, I. 1951. A survey of Indian River Archeology, Florida. Yale University Publications in Anthropology, No. **44**. New Haven, Conn.

SCHOFIELD, J. C. 1960. Sea level fluctuations during the last 4000 years as recorded by a Chenier Plain, Firth of Thames, New Zealand. New Zealand. Geol. Geophys. **3**(3): 467–485.

SOPER, J. H. 1956. Some families of restricted range in the Carolinian flora of Canada. Trans. Roy. Can. Inst. **XXXI,** Pts. II. : 67–90.

WEDEL, W. S. 1941. Environment and native subsistence economies in the central Great Plains. Smithsonian Miscellaneous Collections. **101**(3): 1–29.

WEDEL, W. S. 1953. Some aspects of human ecology in the Central Plains. Am. Anthropologist. **55**(4): 499–514.

LATE PLEISTOCENE SOIL DEVELOPMENT, GLACIATION, AND CULTURAL CHANGE IN THE EASTERN MEDITERRANEAN REGION

H. E. Wright, Jr.
Department of Geology, University of Minnesota, Minneapolis, Minn.

Introduction

Current theories of Pleistocene climatic change are centered on explaining the phenomena of the glaciated and periglacial regions, where glacial deposits, periglacial loess and frost features, buried soils, and fossil-bearing sediments give ample evidence for the nature and sequence of past climates. For Europe it is generally presumed that the atmospheric circulation patterns were so modified that storm tracks shifted south of the Alps to the Mediterrenean region, which experienced a pluvial climate but, at the same time, was sufficiently mild to serve as a refuge for the central European flora and fauna. It is the province of the Mediterranean geologist and paleontologist, however, to locate the field evidence to confirm or deny this presumption, and to work out the correlation of paleoclimatic events between the two regions.

For the eastern Mediterranean hinterland the task is somewhat more important because the region is farther removed from possible secondary effects of the north European and Alpine ice sheets, which together had a profound effect on the climate and circulation patterns of central Europe. Furthermore, the great mass of the Asiatic continent has its own effect on the circulation of this region. It therefore cannot be assumed that the climatic chronology of this region matches that of Europe in its main details.

For the eastern Mediterranean coastal area we have available an independent means of connection between middle-latitude glaciation and low-latitude climatic changes in the form of marine terraces and associated sand dunes. These features record fluctuations in sea level caused by glaciation elsewhere in the world. It might be hoped that the coastal features are traceable inland along rivers to the hinterland, but unfortunately the topography and hydrography of the eastern Mediterranean region are such that the inland tracing can be extended only a few miles.

In the present paper some of the evidence of Pleistocene climatic changes in the eastern Mediterranean will be critically reviewed, and conclusions will be offered concerning the relations between climatic changes and cultural evolution in this region.*

Buried Soils

In the glaciated or periglacial regions, buried soils indicate interglacial or perhaps interstadial intervals marked by ice retreat, stabilization of the terrain by vegetation, and prolonged chemical weathering. The widely accepted

* The field work on which this discussion is based was aided by grants from the Oriental Institute of the University of Chicago, Chicago, Ill.; from the Guggenheim Foundation, New York, N.Y.; and from the Wenner-Gren Foundation for Anthropological Research, Inc., New York, N.Y. For the most part the writer was attached to the projects in prehistoric archeology directed by Robert J. Braidwood of the Oriental Institute.

climatic curve for the Würm glacial stage in Europe is based largely on the sequence of loesses and buried soils in Lower Austria and adjacent Czechoslovakia (Gross, 1958). In the Mediterranean region, however, where there was no lowland glaciation or extensive frost action and loess deposition, buried soils are not so easy to interpret. Two examples on the coast of Lebanon may be described.

South of the Beirut peninsula is a large area of cemented sand dunes that

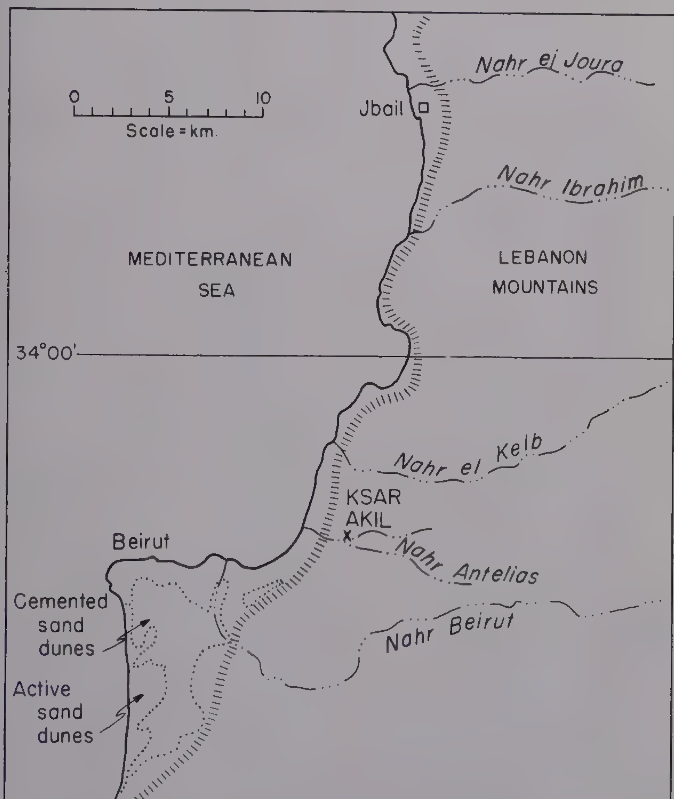


FIGURE 1. Map of Lebanon coast showing location of sand dunes south of Beirut and location of the Paleolithic rock shelter of Ksar 'Akil. Scarp symbol shows steep front of Lebanon Mountains.

have been quarried for building stone for the city (FIGURE 1). In the quarries as well as the sea cliffs the cross-bedded sandstone is interrupted by reddened layers a few feet thick that represent intervals of soil formation (FIGURE 2). The blowing sand was apparently stabilized by vegetation, the calcareous sand grains partially leached, and the iron oxidized to a red-brown color. The leaching solutions precipitated the carbonate in the sand beneath the soil, thereby converting the sand into a firm rock (Wright, 1961*b*). The main buried soil, near the top of the section, is marked by long conical pendants that are terra rossa in limestone or other calcareous material.

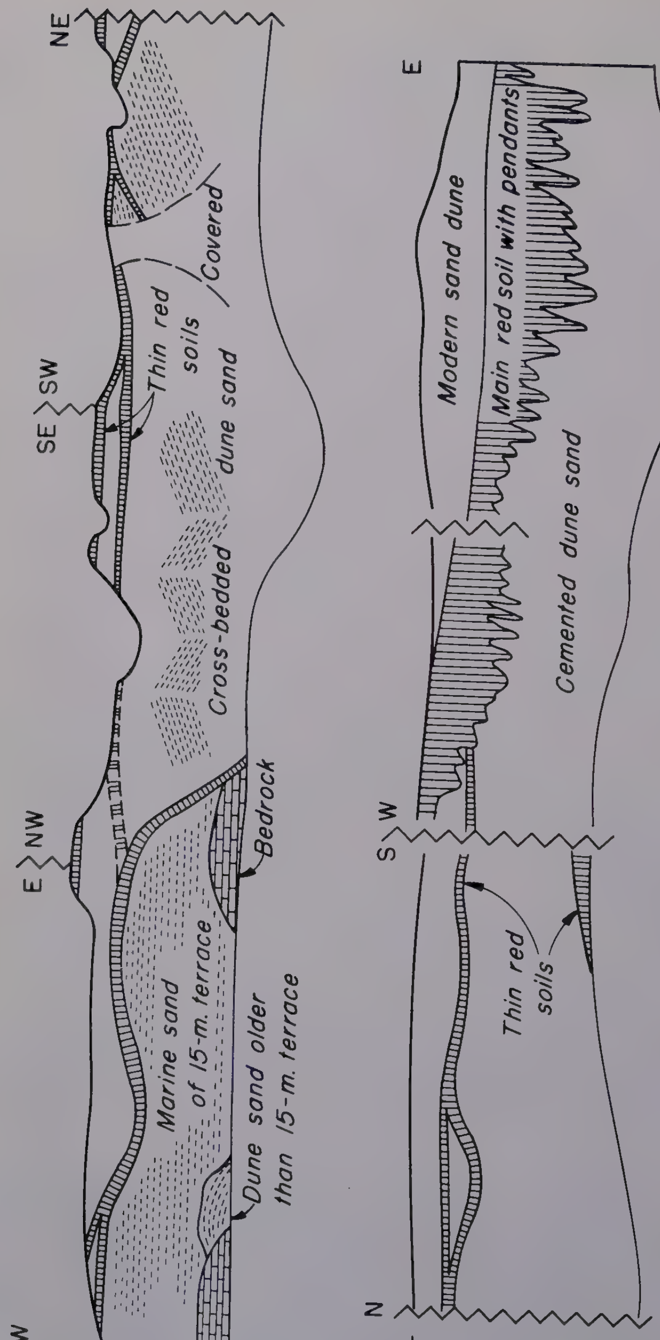


FIGURE 2. Sketch cross sections through late Pleistocene cemented sand dunes and red soils south of Beirut, as exposed in sea cliffs and in rock quarries.

The cemented dunes rest in part on the marine sands or the bedrock terrace that represent the last-interglacial high stand of sea level about 15 m. above the present. The dunes are now being cliffed by the sea, and they clearly extend below present sea level. They were formed at a time of low sea level when beach sands were newly exposed to wind erosion. The correlation of the dunes accordingly is with the low stand of sea level during the last glacio-pluvial epoch of the Pleistocene.

The buried red soils within the cemented sand dunes present a problem in climatic interpretation. Although the soils clearly represent times of cessation of wind deposition in the area, several factors may control the eolian activity in coastal regions. Sand dunes are formed when there is a sufficient supply of sand, winds strong enough and persistent enough to transport the sand the requisite distance, a topography suitable for the deposition of dunes, and vegetation that is so sparse that the sand is not excessively impeded in its movement. The aggressiveness of the vegetation may be controlled by the climatic relations, and also, perhaps, the nature of the dune-forming winds may be climatically controlled. The source of the sand, on the other hand, may be determined by geologic factors unrelated to the climate, such as the bedrock configuration of the coast or the activity of coastal streams that may supply alluvial sand to the coast by longshore transport. In the contest between sand transport by the wind and sand stabilization by vegetation it may be visualized that stabilization (and thus soil formation) may win out only because of a change in the activity of the coastal stream or the longshore current or a change in the configuration of the coastline as the sea regressed during the last glacial epoch. Of course a change in the behavior of a coastal stream may be climatically induced, but the activity of the dunes themselves must not be assumed to indicate the character of the climate.

It may readily be seen that the relations are complex, and the correlation of three buried red soils in the Beirut quarries (FIGURE 2) with three stadial or interstadial phases of the Würm is pure speculation. It can be stated only that the principal soil, which contains the deep pendants and is buried only by weakly cemented dune sand, contains flint artifacts ranging from Levallois-Mousterian to Eneolithic (Fleisch, 1956). The older artifacts were probably dropped in the sand during the time of dune accumulation, and have been concentrated in the soil by the subsequent leaching action that lowered the soil surface. The younger artifacts, however, were dropped during the time of soil formation, and there is a greater accumulation on the surface of the buried soil than within the soil itself. This main soil then may represent primarily the time of the late-glacial and postglacial rise in sea level (the Flandrian transgression), when sand supply was reduced by the steady transgression of the shoreline. The earlier buried soils represent temporary times of stabilization that may or may not be related to climatic changes or sea-level changes.

A second type of buried soil on the coast of Lebanon is found in the Paleolithic rock shelter of Ksar 'Akil, 6 km. north of Beirut. Here about 23 m. of archeological debris accumulated during the Upper Levallois-Mousterian and Upper Paleolithic cultural stages at the base of a tall overhanging limestone cliff facing south on the side of a small coastal stream (Wright, 1951). The lower 7 m. of debris contains also some alluvium deposited as the stream

periodically migrated over to the cliff, but the remaining 16 m. of material apparently built up more rapidly than the valley floor, and thus the stream was kept away from the door (FIGURE 3).

The alluviation of the stream is considered to represent the last glaciopluvial stage of the Pleistocene (Würm), because a similar coastal stream 50 km. to the north (Nahr ej Joura, near Jbail), contains a comparable alluvial fill that interfingers along the coast with cemented sand dunes of the same type as those described south of the Beirut peninsula. These dunes and alluvium rest on the 15-m. marine terrace of the last interglacial stage; they extend below present

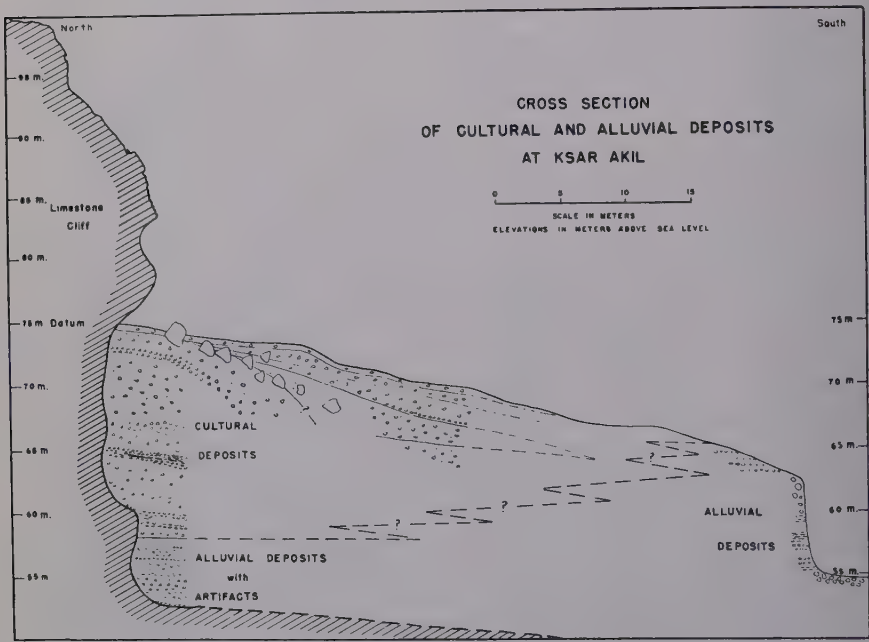


FIGURE 3. Stratigraphic section through the archeologic deposits of Ksar 'Akil, showing relation to alluvial terrace. Buried red soil and stone layer at 65 m. (10.5 m. below surface of deposit). Reproduced by permission of the *Journal of Near Eastern Studies*. (Wright, 1951, v. 10, p. 113.)

sea level and are accordingly correlated with the Würm glacial stage of low sea level.

The filling at Ksar 'Akil is dominantly typical light brown "cave earth," consisting of angular limestone fragments from the roof, pieces of flint and bone, charcoal, and snail shells, all in a matrix of silt and sand tracked in or blown in from outside the shelter. At a depth of 10.5 m. in the deposit, however, there is a distinct layer of red clay about 30 cm. thick overlain by a layer of angular stones of equal thickness. The red clay has the appearance of the typical terra rossa developed on limestone on the Jurassic bedrock of the region, and analysis of the entire stratigraphic section of the rock shelter shows that this

layer has a much lower content of carbonate and higher content of iron oxide and residual quartz compared to the average brown earth of the deposit. The red clay is considered to be a soil developed *in situ* in the rock shelter.

The paleoclimatic significance of the buried soil is difficult to evaluate. Red clay soil is widespread on the limestone of the region, and appears to be forming at the present time. It must be admitted, however, that essentially no red soil has formed in the last 10,000 + years on top of the Upper Paleolithic deposits of Ksar 'Akil, although minor disturbance of the surface by people and animals may have destroyed some incipient soils. On this basis the soil at 10.5 m. might be considered some phase in the Pleistocene when weathering of limestone occurred at a more rapid rate than at present, presumably a pluvial phase.

On the other hand the buried soil may merely represent a time when the shelter was not inhabited, the red soil normally forming on the surface not being diluted by foreign nonred debris introduced from outside. Although the preliminary study of the artifacts (Ewing, 1947) revealed no conspicuous break in the archeological sequence that might record a time of nonhabitation at the 10.5-m. level, more quantitative studies of typology and frequency of the worked flints should provide the information necessary to evaluate this hypothesis. Radiocarbon dating at Ksar 'Akil indicates that the red soil was formed during some interval between 44,400 and 28,500 years ago, so the time of soil formation falls in the time of the early or middle Würm according to the chronology of Europe (Gross, 1958, p. 181; Wright, 1961*a*).

In the glaciated and periglacial regions of Europe buried soils are among the most important indicators of interglacial or interstadial conditions because during the glacial phases this region was either covered by ice or was marked by tundra vegetation and a frigid climate without formation of zonal soils, as indicated by pollen-analytical studies and by the distribution of frost soils. Detailed studies of the soils morphology and chemistry provide information about the nature of the vegetation cover (whether forest or grassland); thus the climate and the thickness and degree of development of the soil horizons in comparison with postglacial soils provide a means of estimating the length of time required for the formation of the buried soil.

In the Mediterranean region the problem is complicated by the fact that the Pleistocene climate, at least in the relatively more humid areas such as Lebanon, was suitable for some kind of soil formation in the interpluvial as well as during the pluvial phases, so the mere presence of a buried soil does not require its correlation with a pluvial phase unless it can be demonstrated that the lack of soils during intervening intervals was a direct result of unfavorable climate. Furthermore, the horizonation of the Mediterranean terra rossa is so weak that it is almost impossible to estimate the level of mean temperature or precipitation or the length of time required for the formation of the soil in question.

In view of these several difficulties, the buried soils of the eastern Mediterranean region—the two examples described above are the best ones known to me—do not provide clear evidence for climatic fluctuations within the Würm glacio-pluvial climatic phase. They certainly do not give the type of quantita-

tive information on mean temperature or precipitation that might help in efforts to reconstruct the Pleistocene general circulation of the atmosphere for these latitudes.

Glaciation

The Lebanon Mountains reach an elevation of more than 3000 m. above sea level and are marked by heavy snowfall and by numerous frost features near their crest (Klaer, 1957). They were not glaciated during the Pleistocene; the great cirquelike basin of the Cedars of Lebanon at 2000 m. was apparently formed by landsliding (Dubertret, 1950) rather than by glaciation.

Farther to the north and east, however, along the Taurus-Zagros mountain arc of Turkey, Iraq, and Iran, Pleistocene glaciers were locally present. Even today a few small cirque glaciers and perennial snow patches exist on sheltered slopes above 3300 m. In the Taurus ranges of southcentral Turkey, Pleistocene cirques are described by Louis (1944) 800 m. below the modern snowline. In the eastern Taurus of the southeastern corner of Turkey, however, north of the Cilo Dagh, small cirques have been identified as much as 1200 m. below the present cirque glaciers and, in the region of the Algurd Dagh of northeastern Iraq, cirques have been found 1800 m. below the presumed snowline of today (Wright, 1960, 1961*c*). Several highland valleys have been greatly broadened by glacial erosion and plugged with morainic debris, which in fact extends down to an elevation of 1100 m. Glaciofluvial deposits as much as 30 m. thick are present in the Ruwandiz River valley downstream from the glaciated area. Certain other valleys, however, even at rather high elevations, are devoid of glacial features, because the exposure of the snow-collecting basin or cirque with relation to the warm summer sun is apparently critical in the development of these glaciers. The greatest concentration of features is in the valleys leading out of the north slope of the Algurd Dagh in Iraq and out of the Cilo Dagh and Kara Dagh in Turkey.

Farther southeast along the Zagros Mountains in southwestern Iran, as well as in the Iranian Plateau northwest of the ranges, Pleistocene glaciers were either absent or were confined to much higher elevations.

The present climate of the Tauros-Zagros mountain arc is basically Mediterranean, with winter storms and summer drought. Much of the precipitation in the mountains is snow brought by cold fronts associated with cyclonic disturbances that either have traveled the length of the Mediterranean or have regenerated as the shallow Cyprus lows in the eastern Mediterranean region (El Fandy, 1946; Butzer, 1958, p. 22). These disturbances are strong and stormy, and some of them traverse Mesopotamia and reach the Persian Gulf and India, especially in the spring. During the winter the penetration is inhibited by the development of the shallow Arabian anticyclone, but this air mass provides further precipitation in the Kurdish mountains in the form of relatively warm and gentle storms of the orographic type as the air is forced northward up the mountain flanks (Boesch, 1941).

The maximum precipitation in the entire Taurus-Zagros arc is in the Kurdish mountains of the southeastern corner of Turkey and in adjacent Iraq (FIGURE 4). The crests are higher and broader in this region than elsewhere, exceeding

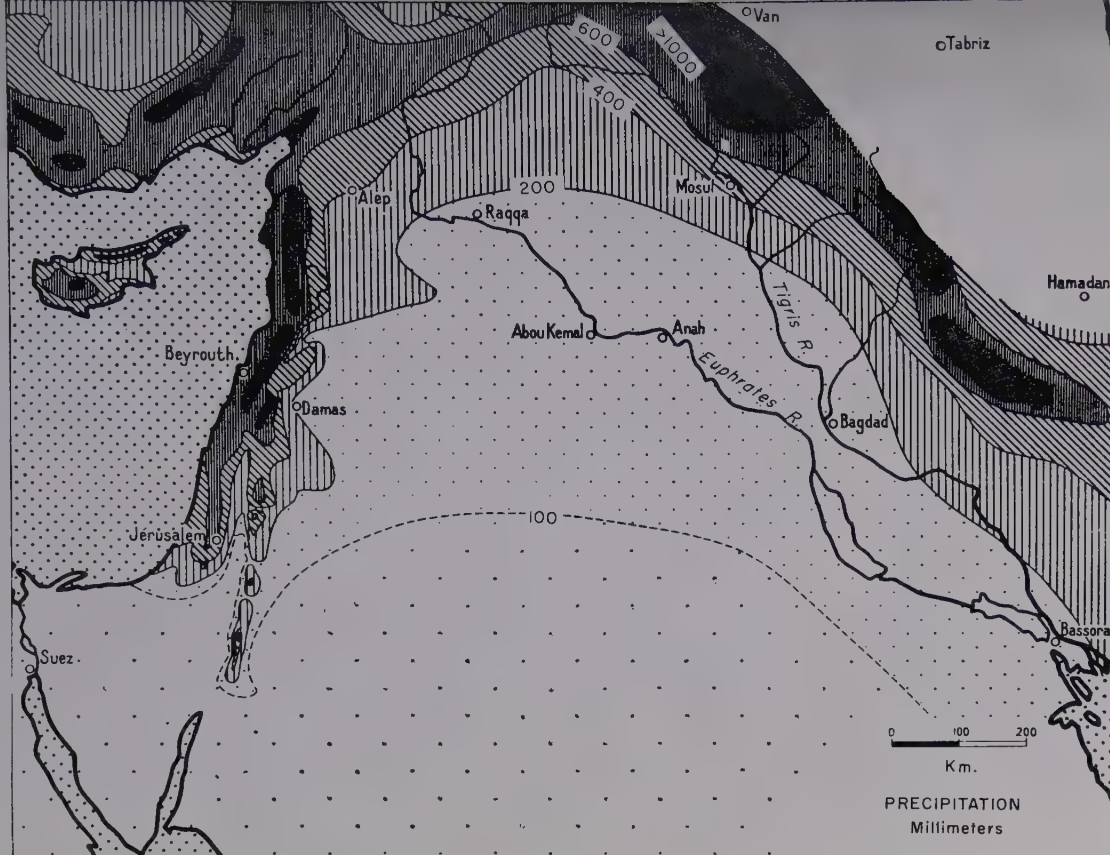


FIGURE 4. Precipitation map of the eastern Mediterranean region showing control of precipitation by the Lebanon Mountains and by the Taurus-Zagros mountain arc. The Algurd Dagh and Cilo Dagh area of Pleistocene glaciation lies northeast of Mosul. Map adapted from Fish and Dubertret (1946).

3000 m. over extensive areas and reaching summits well over 4000 m. They thus occasion greater uplift of the invading moist air masses. Also, the trend of the ranges changes here from northeast to southwest. Most of the cyclonic storms enter the region from the Mediterranean across the "Syrian Saddle" with a trajectory to the northeast along the front of the western and central Taurus ranges. Where the mountain arc turns to the southeast these storms may be forced to rise high over the mountains or be deflected to the southeast.

Life Zones and Early Man in the Taurus-Zagros Mountains

The climatic relations are reflected by the vegetation zones. Close to the northeast corner of the Mediterranean Sea, where the precipitation is relatively high and the winters mild, the forest grows down to sea level and its upper limit is at an elevation of about 2000 m. Eastward the lower forest limit rises because of increasing aridity at low elevations and, in northwestern Iraq, it has an elevation of about 700 m. The forest or, more properly, woodland, consists primarily of oak. Some pine occurs close to the Mediterranean, and juniper extends east along the mountains, generally at high elevations. Walnut, maple, hawthorne, almond, and pistachio occur in minor amounts. The forest is locally thinned, reduced to scrub, or even eliminated by woodcutters and goats, especially near its lower margin and close to the more heavily populated or traveled areas. It generally requires an annual precipitation of at least 500 mm. It is succeeded below by steppe: the domain of intensive dry farming and grazing. At still lower elevations, where the precipitation is less than 200 mm., the steppe is succeeded by desert steppe and, in turn, by desert.

The forest limits of 2000 m. and 700 m. continue southeastward along the outer flank of the Taurus-Zagros ranges to the end of the mountains in southern Iran. On the plateau side of the mountain belt these elevation limits do not apply because of the low precipitation. The tree cover barely extends into the plateau despite the general elevation of 1200–1700 m., and the succession of steppe to desert steppe is controlled more by distance northward from the mountains than by elevation.

The Pleistocene depression of snowline in the Zagros-Taurus Mountains as inferred from the cirque observations must have been accompanied by depression of the life zones as well. In the Alps it has been found through pollen analysis that the upper treeline was in fact depressed more than the snowline (Firbas, 1939). Depression of the snowline as much as 1800 m. in the Zagros Mountains implies a temperature depression of 9° C. if a vertical temperature gradient of 0.5° C./100 m. is assumed. In fact, temperature measurements at a series of ground stations extending from the Persian Gulf across the Zagros Mountains to the Iranian Plateau show that the present lapse rate (both mean annual and mean July) is closer to 0.7° C./100 m., implying a temperature depression for the Pleistocene of at least 12° C. (Wright, 1961*c*). Such a figure seems excessively large when it is applied to the Mesopotamian piedmont and lowland, for it implies that the vegetation zones were shifted downward from their present elevation ranges so that most of the mountains were above the treeline and most of Mesopotamia was forested. It seems more likely that a

modest depression in temperature was combined with an appreciable increase in snowfall in the mountains under consideration so that the glaciers could form at the relatively low levels indicated. The special location, trend, and height of these mountains with respect to more frequent storm tracks from the Mediterranean might explain the postulated great increase in snowfall.

The inferred shift in life zones in the Zagros Mountains during the last glacial period must be tested by paleontologic studies. At the moment we have only the opinion of Reed (Reed and Braidwood, 1960, p. 169) that the mammalian and molluscan faunas from Pleistocene Paleolithic caves in this region suggest a vegetation and, thus, a climate similar to today's. Because of the fact that many of the animals represented may have been hunted in other vegetational zones and imported to the caves, however, it is necessary to provide other types of paleontological evidence. Accordingly pollen-analytical and paleolimnological studies have been started at several localities in the Zagros Mountains.

The effects of these climatic and vegetational changes on the evolution of early man are uncertain. As a result of the excavations of Paleolithic and later materials by Solecki (1955) at Shanidar and of Braidwood *et al.* (1960) at numerous other sites in northeastern Iraq, the sequence of cultural changes from cave to village is becoming better known, both in the nature of the material objects and—thanks to radiocarbon dating—in the chronology.

Mousterian cave sites are numerous in the region, and perhaps the largest, at Shanidar, shows a significant unconformity between the Mousterian and Upper Paleolithic layers that has been dated as marking the time between about 25,000 and 12,000 years ago. Solecki (1955, and personal conversation) has made the suggestion that this time represents the interval of maximum glaciation and depression of the life zones in the mountains and the consequent abandonment of the cave, which is surrounded at present by oak forest. This correlation fits what we know about the chronology of glaciation in the Alps, where the ice reached its maximum about 20,000 years ago and had retreated well up into the mountains by 12,000 years ago.

The first open living sites in the region were apparently first occupied shortly after 12,000 years ago and, by 9000 years ago, full-scale villages such as Jarmo had been established, having domesticated animals and grains. The problem is posed whether or not the cultural changes associated with the establishment of agriculture and villages were impelled by the climatic changes. For a definite answer to this question we must await more precise knowledge about the vegetation zones during this critical period—and we may hope that pollen analysis may furnish this information—but at present it would seem that the most important climatic change that closed the Pleistocene had been completed by this time. If the interpretation of the Shanidar unconformity is correct, we can say that the cave and the region were habitable again 12,000 years ago. Regardless of the timing, however, it is clear that the location and gross topography of the mountains and foothills were such that climatic changes must merely have shifted the life zones up and down. It is probable that a suitable environment for food collecting and, ultimately, food cultivation existed over a broad area, at least in the latest Pleistocene and early Recent.

References

Braidwood, R. J. & B. Howe. 1960. Prehistoric investigations in Iraqi Kurdistan. Oriental Inst. Univ. Chicago. Studies in Ancient Oriental Civilization, No. **31:** 184 p.

Boesch, H. H. 1941. Das Klima des Nähen Ostens. Naturforsch. Ges. Zürich, Vierteljahresschr. **86:** 8–61.

Butzer, K. W. 1958. Quaternary stratigraphy and climate in the Near East. Bonner Geogr. Abhandl. **24:** 157 p.

Dubertret, L. 1950. Carte geologique au 50,000. Feuille de Baalbek. Republic of Lebanon, Ministry of Public Works. Beirut, Lebanon.

El Fandy, M. G. 1946. Barometric lows of Cyprus. Roy. Meteorol. Soc. Quart. J. **72:** 291–306.

Ewing, J. F. 1947. Preliminary note on the excavations at the Paleolithic site of Ksar 'Akil, Republic of Lebanon. Antiquity. **21:** 186–196.

Firbas, F. 1939. Vegetationsentwicklung und Klimawandel in der mitteleuropäischen Spät- und Nacheiszeit. Naturwissenschaften. **27:** 81–89, 104–108.

Fleisch, H. 1956. Dépôts préhistoriques de la côte libanaise et leur place dans la chronologie basée sur le Quaternaire marin. Quaternaria. **3:** 101–132.

Fish, W. B. & L. Dubertret. 1946. Carte pluviométrique du Moyen Orient. Délégation Generale de France au Levant. Sect. Géol. Notes et Mémoires. **4:** 115–121.

Gross, H. 1958. Die bisherigen Ergebnisse von C-14 Messungen und paläolithischen Untersuchungen für die Gliederung und Chronologie des Jungpleistozäns in Mittel-Europa und den Nachbargebiete. Eiszeitalter und Gegenwart. **9:** 155–187.

Klaer, W. 1957. Beobachtungen zur rezenten Schnee- und Strukturbodengrenze im Hochlibanon. Z. Geomorphol. **1:** 57–70.

Louis, H. 1944. Die Spuren eiszeitlicher Vergletscherung in Anatolien. Geol. Rund. **34:** 447–481.

Reed, C. A. & R. J. Braidwood. 1960. Toward the reconstruction of the environmental sequence of northeastern Iraq. Oriental Inst. Univ. Chicago. Studies in Ancient Oriental Civilization. **31:** 163–174.

Solecki, R. S. 1955. Shanidar Cave, a Paleolithic site in northern Iraq. Smithsonian Inst. Ann. Rept. **1954:** 389–425.

Wright, H. E., Jr. 1951. Geologic setting of Ksar 'Akil, a Paleolithic site in Lebanon—preliminary report. Near Eastern Studies. **10:** 115–119.

Wright, H. E., Jr. 1960. Climate and prehistoric man in the eastern Mediterranean. Oriental Inst. Univ. Chicago. Studies in Ancient Oriental Civilization, No. **31:** 71–98.

Wright, H. E., Jr. 1961*a*. Late Pleistocene climate of Europe—a review. Geol. Soc. Am. Bull. **72:** 833–883.

Wright, H. E., Jr. 1961*b*. Late Pleistocene geology of coastal Lebanon. Quaternaria. In press.

Wright, H. E., Jr. 1961*c*. Pleistocene glaciation in Kurdistan. Eiszeitalter und Gegenwart. In press.

PALAEOCLIMATOLOGY AND ARCHAEOLOGY IN THE NEAR EAST*

Ralph S. Solecki
Columbia University, New York, N.Y.

Arlette Leroi-Gourhan
Musée de l'Homme, Paris, France

Archaeological investigations combined with interdisciplinary studies in allied fields at Shanidar Cave and the nearby village site of Zawi Chemi Shanidar in the Zagros Mountains of northern Iraq have produced data that enable us to establish climatic conditions there on a relatively firm basis from at least 50,000 years ago.

The cave, reported by Solecki (1952, 1953*a* and *b*; 1955*a* and *b*; 1959, 1960), and Solecki and Meyer Rubin (1958), situated at 36° 50′ N lat., and 44° 20′ E long. is one of the deepest stratified archaeological sites in the Near East. Botanical remains have been found and identified which, coupled with a series of 16 carbon-14 dates extending back 50,000 years, gives us an unusually good index of climatic changes there. The carbon-14 samples were check dated by four separate laboratories: Columbia University's Lamont Geological Observatory at Palisades, N.Y., the United States Geological Survey laboratory in Washington, D.C., the University of London Geochronological Laboratory, London, England, and the Groningen Laboratory, University of Groningen, Groningen, the Netherlands. The dated layers of the cave extend down to about the upper third of the deposits. The advantages of a dated sequence of a unit of cave deposits are readily apparent. There is no fear of losing the thread of continuity as in a reconstructed sequence from several sites.

One primarily archaeological and two primarily geologically based reconstructions of the Near Eastern climatic sequence have been recently attempted. F. Clark Howell (1959) has drawn up a prehistory of the Levant based essentially upon a typology of the stone industries and, to some extent, on the soil characteristics of the sites. Karl W. Butzer (1958) has worked with geological considerations, and some dependence on archaeological sequences. The more sophisticated analysis of Herbert E. Wright, Jr. (in R. J. Braidwood and B. Howe, 1960) is based principally upon geological evidences. Howell's work is open to criticism since there is doubt that similar archaeological (or homotaxial) industries are necessarily contemporaneous in wide geographic areas and, in the Near East, the situation is complicated further by differing environments. Furthermore, the stratigraphic framework of his sequences, based as it is on the works of excavators of different backgrounds and training, is also in part open to question.

Of the two geologically based studies, Wright has evaluated the data more critically. Karl Butzer did not have certain of the literature available to him. Wright has used the geological data as well as his own original work in the Near East for his reconstruction. However, it is apparent that geological findings of

* The work described in this article was supported in part by grant from the National Science Foundation, Washington, D.C.

the kind used by these investigators give only gross temporal implications and are not too satisfactory for finer separations in cultural deposits. In fact, Wright's colleagues, Charles A. Reed and R. J. Braidwood (in R. J. Braidwood and B. Howe, 1960, p. 163) could not confirm the evidence for climatic changes as postulated by Wright.

Taking the cue from palaeobotanical research in other areas such as, for example, northern Europe, where these studies are far advanced, investigators in Near Eastern palaeoclimatology have recognized that the most sensitive index of climate obtainable is that constructed from pollen analysis. Thus far no such sequence has been available in the Near East. There are certain difficulties attendant upon pioneer work of this sort in this region. As Wright has pointed out (in R. J. Braidwood and B. Howe, 1960, p. 96), the flora of the eastern Mediterranean area is not well known, and there are few good lake and swamp sites from which a climatic sequence could be obtained. Dated material from such sites is another factor for consideration.

In default of pollen remains, the palaeoclimatological sequence therefore had to depend upon such biological and geological evidences as faunal remains, mollusks, sea levels where applicable, terraces, soil analysis, and other indicators. Faunal remains and mollusks have played a great part in attempts at gauging climatic sequences in archaeological sites.

Regarding faunal remains, we recognize that cultural selection, or the cultural filter through which the material passes, plays a large part in the presence or absence of remains in archaeological deposits. We do not have to dig for such evidence. Today, for instance, wild boars run free in Moslem Kurdistan, Iraq, whereas the other wild animals are almost extinct. No wild boar remains can be found in the upper modern layers at Shanidar Cave nor at the village site of Zawi Chemi Shanidar.

Man is disposed to hunt one type of animal in favor of another for other more obvious reasons. For instance, in archaeological deposits the proportion of herbivore animal remains represented are overwhelming as compared with the carnivore remains. The simple reason for this is that the latter are more difficult and dangerous to hunt and are also probably less numerous (a factor undoubtedly not so important as the first two). In short, there are a number of very excellent reasons why certain animal species are heavily represented in the archaeological record, and others not so numerous.

The classic climatic sequence based upon faunal remains is the Mt. Carmel, Israel, sequence (D. M. A. Bate, in D. A. E. Garrod and D. M. A. Bate, 1937; F. E. Zeuner, 1958) has received a number of criticisms, some well grounded and others, perhaps, less justifiable (M. Stekelis and G. Haas, 1952; R. Vaufrey, 1938, 1939; F. C. Howell, 1959). The criticism ranged from the method of collection of the faunal remains (the basis of the hypothesis) to the cultural factors involved. Most recently, the sequence has gone the full circle, and has received acceptance again (C. B. M. McBurney, 1960, pp. 47–48).

Mollusks have offered similar difficulties. Although they occur in the area of Shanidar and have been found in the deepest deposits of the cave, they were certainly not native to the cave interior. They must have been brought in either by whim of man, accident, or caprice of nature, or for food. The only

horizon in which they occur in great numbers is in the Mesolithic and Proto-Neolithic layers. At that time, for some reason, people about 10,500 to 12,000 years ago relished snails, and appeared to have eaten great quantities of them. Snails furthermore, are not so precise indicators of climate as it was once hoped they would be (C. A. Reed and R. J. Braidwood, in R. J. Braidwood and B. Howe, 1960, p. 164); yet Reed and Braidwood (ibid, p. 165) use snail shells as a final climate gauge on the open site, Barda Balka, in Iraq.

We note concerning the environmental sequence of northeastern Iraq that Reed and Braidwood (in R. J. Braidwood and B. Howe, 1960, p. 163) have the impression, based on the studies of animal bones, snails, and certain of the woody floral remains that there has been little change in the biota of that region in late glacial and postglacial time. This, despite Wright's evidence for mountain glaciation, terrace building, and alluviation, points to climatic shifts. Our own climatological data from Shanidar Cave, based on palaeobotanical evidence and trace element studies have shown that there are truly marked fluctuations in climate in this region. In fairness to Reed and Braidwood however, they have indicated that their observations might be subject to revision. It is evident that geological and botanical data not culturally induced are more useful criteria for climatological purposes than the materials of archaeological deposits which are laid down through a cultural filter. Man could not control the natural phenomena such as floods and the dispersal of pollens. Nevertheless reflection on the subject of pollen deposition shows that even palaeobotanical evidence should be accepted with some caution. The general cautions to be considered are again the natural and unnatural agents. The former include the elements, such as wind and water. The latter include man as the primary agent. In the front of caves, the wind is the principal carrying agent of pollens. On the other hand, toward the rear of the caves, in the corridors and recesses, where the wind has less access, the principal carrying agent may be man or animals: the pollen being brought in on the feet or in the furs. It is quite conceivable that man brought pollens into the site or cave with his food; that is, the pollens may have been trapped in the fur of the newly captured or killed animal, or in such vegetal food as he brought in: even with the boughs and foliage and his very bedding of spread grasses. This would certainly change the distribution of pollen remains in cave sites or other habitations. Furthermore, we are apprised of the fact that, on the basis of modern sampling of pollen, what is found in the deposits, whether it is a lake deposit or an occupational cave deposit, may not necessarily represent the true assemblage and percentage of floral remains in the area.

Anticipating future cave deposit pollen studies, we can be sure that the correlating of their pollen diagrams may not be easy. No two caves have the same geographical and ecological environments; moreover no two caves will have the same cultural deposits.

The Prehistory of Shanidar Cave and Zawi Chemi Shanidar

These two sites are situated near each other, the cave site on the side of Baradost Mountain, and the village site (Zawi Chemi Shanidar) on a terrace above the Greater Zab River. The deposits of the cave site, nearly 14 m. deep to bedrock, have been divided into four major cultural layers, lettered in

FIGURE 1 from top to bottom, A, B, C, and D. The first layer (A), has been identified as from Modern to Neolithic in age; the second layer (B) has been subdivided into B1 and B2 with the corresponding cultural tags Proto-Neolithic and Mesolithic; the third layer (C) is called Upper Palaeolithic or, here,

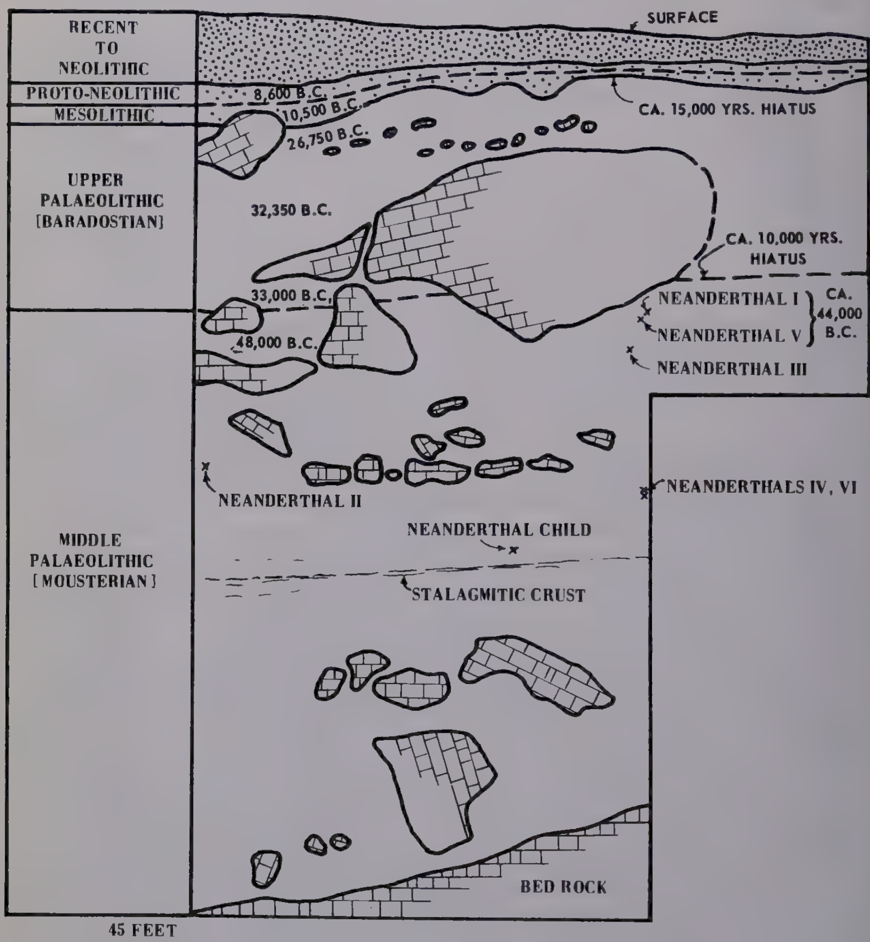


FIGURE 1.

Baradostian; and the fourth layer, the thickest, is the Middle Palaeolithic Mousterian layer. There is a marked chronological and cultural hiatus between layers A and B, between B and C, and between C and D.

The village site of Zawi Chemi Shanidar corresponds culturally and chronologically with layer B1 of Shanidar Cave. In fact, it would appear that the village people wintered in the cave and summered on the river terrace.

The Pollen Analysis

Initially Gunnar Erdtman of the Palynological Laboratory of the University of Stockholm, Stockholm, Sweden played a role in the palaeobotanical studies. He had collected a series of samples himself in the field at Shanidar in 1957. However, he reported that his findings proved to be pollen-sterile.

In the event that the lot shipped to Erdtman should go astray or suffer mishap, a duplicate series of the soils was collected. Following the news that the samples examined by Erdtman were sterile (using his method), Solecki learned that Arlette Leroi-Gourhan of the Musée de l'Homme was interested in trying her method of pollen analysis on the duplicate Shanidar samples. Seven soil samples from the Zawi Chemi Shanidar village site and 23 soil samples from Shanidar Cave were sent to her through the Smithsonian Institution, with which Solecki was then associated.

Leroi-Gourhan's method of pollen analysis has yielded the following results, as stated in her preliminary announcement.* Thus far she has been able to get five determinations in layer D, two in layer C, and two in layer B of Shanidar Cave. In the village site, she has found three determinations. These determinations, coupled with the carbon-14 dates available for Shanidar, give us an assessment of the palaeoclimate in this part of the Near East. This should not be construed to mean that the total picture is clear; however, since it is certainly evident that there are gaps in the present chain of evidence that need filling. It is hoped that some of these links may be forthcoming in the new series of samples sent to Leroi-Gourhan from the 1960 Shanidar season.

The data incorporated in Leroi-Gourhan's preliminary study is submitted below. "Datum" measurement is at or about the surface level of the cave floor.

According to Leroi-Gourhan, the study of fossil pollen grains collected in cave sites differs greatly from that of bog samples. Since there are only a few, poorly known analyses of cave samples, it seems useful to present a few preliminary remarks on these differences.

Due to favorable conditions of preservation, particularly the humidity, the number of pollen grains found in bog samples is very high. A thousand pollen grains or more are available on one slide. The study can be completed relatively rapidly and is based on a valid statistical sample. In cave samples, on the contrary, the number of pollen grains on each slide ranges usually from only 1 to 15. This makes it necessary to examine numerous slides and, in certain cases, even to prepare 2 or 3 samples from the same archaeological layer in order to obtain at least 30 to preferably 100 pollen grains for study. In some rare cases, 2 or 3 thousand pollen grains are available, but this is exceptional.

Pollen grains found in bogs for the main part had been deposited by the wind. From the botanical point of view this is suspect since the pollen spectra are not truly representative of the local vegetation. Anemophilous pollen grains such

* "First Data for the Climatological Study of Shanidar Cave, Iraq." Translated by Jacques Bordaz, Research Associate, Columbia University. R. Solecki is responsible for any errors committed in the use of Leroi-Gourhan's data, and for the material other than that submitted by her.

as those from pines and walnut trees predominate in the samples; furthermore, they might have traveled long distances.

From this point of view, caves offer certain advantages. The pollen grains deposited in them came from the immediate vicinity, picked up in nearby bushes and grasses by animals and man. The pollen grains could be brought into the cave on the feet and clothing of people and on the furs of animals. The relative proportion of botanical species would therefore be relatively better preserved, and the pollen spectra would represent vegetation existing in the immediate vicinity of the cave. An open-air settlement such as Zawi Chemi Shanidar is closer palynologically to a cave habitation site than to a peat bog. The small differences existing between the samples of Zawi Chemi Shanidar and those of the corresponding layers of Shanidar Cave are probably due to their own setting and to a small but distinct time difference.

Pollen Analysis of the Shanidar Cave Soils

The Middle Palaeolithic (Mousterian) layer D. In the cave, the first sample (8.6 m. below datum) was collected in the upper middle part of the Mousterian layer in a stalagmitic crust indicative of a wet period. A total of 121 pollen grains from this layer have been studied, 35 of which are pollen grains of *Phoenix dactylitera* (date palm).

The flora is quite diversified and probably includes pollen grains from different vegetation zones existing on the slopes leading to the cave. Illustrating this possibility among the pollen are two pollen grains of *Quercus* (oak), one of which is of *Quercus ilex*; the other one is probably of a deciduous species. It is unlikely that these two trees grew side by side. The date-palm trees were probably located at the bottom of the valley, about 1000 feet below the cave. The following species were probably at higher altitudes along the mountain slopes: *Ulmus* (elm), *Pinus* (pine), *Juglans* (walnut), *Rhamnus* (buck thorn). Evidence for a somewhat dry, grassy flora is also found. Most of these pollen grains are Cichoriaceae (27 pollen grains), Chenopodiaceae (10 pollen grains) and Labiatae.

The fossilization of the pollen grains in this sample is unusual. The external layer or exine has become transparent, although probably very hard. For instance, certain Cichoriaceae, easily recognizable by their pores and spines, look very much like crystals.

The second sample (7.5 m. below datum) to be studied was collected from the occupation soil below the skeletal remains of Shanidar Neanderthal II. Four pollen grains of *Abies* (fir) indicate a very significant cooling of the climate. Parts of the mountain slopes were forested and covered with grasses: in other places, dry grasses were still present, including Chenopodiaceae, Anthemideae, and *Ephedra*.

There are too few pollen grains available from the upper middle of the Mousterian layer to permit any valid conclusions to be drawn. Important climatic fluctuations might have occurred in this range of which we do not have any record.

At the top of the Mousterian layer, 2 samples collected at depths of 4.25 m. and 4.35 m. are botanically quite similar and seem to represent the most

typical Shanidar environment of the time. This includes pine, juniper, oak, and chestnut trees on the slopes of the mountain, and date palms and cypress at the bottom of the valley. A trend towards a wetter climate might be more pronounced in the upper part of the Mousterian layer.

The Upper Palaeolithic (Baradostian) layer C. A gap of ca. 10,000 years occurs between the Mousterian and the Baradostian layers. Two analyses from the Baradostian layer were possible. Each indicated a different climatic picture. Near the base of the layer (4.0 m. below datum), the sample indicates the first definitely dry climate that has been recorded at the cave. A total of 229 pollen grains were counted, of which 207 were herbaceous. The pollen spectrum indicates a typical steppe vegetation with Compositae and Labiae, including *Armeria*, *Ephedra*, *Fagopyrum* and *Centaurea*. Present still are a few pollen grains of date palm, although much less than in the lower of the Mousterian layer samples. It cannot be securely established that this very dry period represents an interstadial. Comparing this radiocarbon dated horizon with contemporary climate phases in Europe, it might correspond with the first dry period of the Upper Palaeolithic starting ca. 34,000 B.C. According to radiocarbon dates, a climatic correspondence would therefore seem to have existed at that time between western Europe and northern Iraq.

The sample collected at 3.0 m. depth indicates that wet conditions brought a new increase of forest cover with cold characteristics, well demonstrated by the presence of alder and ash-tree pollens. Furthermore, pollen grains of Graminae are more numerous than those of Compositae.

The Mesolithic (Shanidar B2) and the Proto-Neolithic (Shanidar B1 and Zawi Chemi Shanidar). Material for this horizon at the cave and at the valley site of Zawi Chemi Shanidar may be considered together. The samples are botanically identical, except that trees are more numerous at the mountain site. There is a gap of ca. 14,000 years at Shanidar Cave between the Baradostian layer and the Mesolithic layer, ranging perhaps from 26,000 to 12,000 B.P. This corresponds in Europe with the coldest phase of the Würm, or Würm Maximum. Solecki suggests that the cold conditions are the rather plausible reason why this cave site was abandoned during this interval. Further more Mme. Leroi-Gourhan believes that there is evidence for a flooding of the cave at the end of this period, perhaps due to snow melting. The study of the sample taken at the base of the Mesolithic layer (Shanidar B2) indicates a mixture of two sediments. Pollen grains typical of the period are present, such as those of cypress, pine, and chestnut tree but, in addition, one finds evidence of other much older plants, dating perhaps to the end of the Tertiary. Such plants include ancient ferns, striated pollen grains, and a perfect pollen grain of Acacia. An intrusion of older sediments, probably due to water, seems therefore to be indicated.

Thereafter the climate became drier. The nearness of the village site of Zawi Chemi Shanidar to water is confirmed by the presence of moss spores. However, the pollen spectra generally indicates an increasing condition of dryness.

It is still difficult to draw a large number of inferences on the basis of the Shanidar data since there is a complete lack of comparative material from this region. However, it seems that although a great amount of time must be

devoted to it, the palynology of the Quaternary sediments of the Near East can yield important results.

The study of Bruno E. Sabels, of the Institute of Geophysics, University of California at Los Angeles, Calif., is abstracted below. The arrangements for this work was done by Richard Shutler, Jr., of the Nevada State Museum, Carson City, Nev. Shutler has identified pollen in a Shanidar Cave soil sample that he has examined, although the analysis is not yet completed. The trace element studies of Sabels summarized here have yielded data that on the whole, corroborates Leroi-Gourhan's climatological findings.

Trace Element Study of the Shanidar Cave Soils

An independent study of the soils for pollen and trace elements is being made in the United States, as intimated above. Sabels prepared five samples from Shanidar Cave for emission-spectrographic analysis. It was found that elements of interest in the spectra were Si, Mg, Al, Cu, Na, Ti, Ca, Mn, K, Sr, and Cr. Of particular interest in this group are Cu, Ti, Ni, Mn, Cr, K, Na, and Ca. The samples are described from top downward in the section, as they were recovered.

Modern to Neolithic layer A. A sample taken from 1.4 m. depth indicated in the analysis the elements Na, K, Ca, Cu and Mn. This showed that the climate was similar to the present, with no pronounced aridity. There was no lack of rainfall, yet no pronounced humidity either. The soil type was chestnut.

Proto-Neolithic layer B1. The second sample was collected at the base of this horizon at a depth of 2.4 m. The carbon-14 date for this zone was ca. 10,600 B.P. The analysis (Na, K, Cu, Cr) indicated that the climate was cooler and more moist than the climate indicated in the layer-A sample mentioned above. It was probably subhumid, and is believed to be a transition stage.

Upper Palaeolithic (Baradostian) layer C. The third sample from toward the base of this layer was collected at a depth of 4.85 m. This horizon is given an approximate age based on radiocarbon dating of over 35,000 B.P. The index of climate reflected by all elements is cool and humid conditions. The soil is from chernozem to podzol type.

Middle Palaeolithic (Mousterian) layer D. The fourth sample was collected at a depth of 7.3 m. in a horizon well below the level dated as 50,000 B.P. (carbon-14 sample collected from a depth of 5.0 m.). This sample indicates the coldest and moistest climate for which evidence has been obtained thus far by this method. Most trace-element indicators reach their maximum values (Cr, Mn, Cu, Ti).

The fifth sample, also from this layer, was collected at a depth of 8.3 m. It is the oldest sample examined by Sabels. The trace element analysis of this sample indicated that there was a change of climatic conditions. The climate was probably an interstadial according to Sabels. It represented the climate like that indicated by the sample from layer B or from layer A, although actually neither was as semiarid as the latter nor as subhumid as the former. Sabels

notes that "... the true nature of the last sample ... will become apparent upon analysis of the subsequent material down to the bottom of the cave."

Discussion of Leroi-Gourhan's and Sabels' Findings

The question before us is: How do these findings of Sabels and Leroi-Gourhan correlate with each other? First, however, is the question of how closely in the stratigraphy are the samples matched. Unfortunately Leroi-Gourhan's present tests were made on a different series of samples and from a different set of proveniences than those of Sabels. Moreover the cave deposits do not lie perfectly level, but dip noticeably to the north and west. This gives us the phenomena of a Mousterian age soil sample at 4.35 m. in the eastern part of the excavation (at the top of the Mousterian layer), and an Upper Palaeolithic Baradostian-age sample at a depth of 4.85 m. in the western part of the excavation (at the base of the Baradostian). The samples that Sabels analyzed came from a vertical cut in the north face of the western part of the excavation. Those examined by Leroi-Gourhan were taken from a variety of situations within the cave excavation. However happily, fortuitously it would appear, most of the samples examined by Sabels and Leroi-Gourhan were from closely corresponding horizons and proveniences. Allowing for these discrepancies, there is much agreement between the work of these two independent investigators.

Mention has been made that a number of carbon-14 dates have been obtained from the upper part of the cave deposits. Unfortunately, the carbon-14 dated samples were not always matched by the soil samples for pollen analysis. A date of 50,000 B.P. was taken on a carbon-14 sample from the 5.0 m. level in the west part of layer D. Therefore the soil samples from below this horizon and in the same neighborhood should be older.

Regarding the correlation of the pollen analysis and the trace-element analysis, Sabels' sample from layer D at 8.3 m., which seems to indicate an interstadial to him and a change in climatic conditions, is paralleled by Leroi-Gourhan's 8.6 m. sample from the same layer. This could possibly be the interstadial at about 60,000 to 70,000 years charted by R. F. Flint and F. Brandtner (1961, p. 322) in their paper, "Climatic Changes Since the Last Interglacial." Again in the Mousterian horizon, Leroi-Gourhan's findings from the sample at 7.5 m. compare climatologically with Sabels' interpretations based on trace elements from his 7.3 m. sample. Both observations suggest a very cold and moist period, which may be the stadial of Early Würm in Europe, as indicated on Flint and Brandtner's chart (1961, p. 322).

The sample examined from the 4.35 m. level in the upper part of the Mousterian layer may be dated at about 46,000 B.P. (taken on a sloping layer). This was a rather warm climate in which palm trees evidently grew down in the valley, according to Leroi-Gourhan's analysis. This climate appears to correspond with that of the Göttweig Interstadial in Europe (R. F. Flint and F. Brandtner, 1961, pp. 322, 324).

For the Baradostian layer, we may compare again the interpretations of Sabels and Leroi-Gourhan. The depths and proveniences from which their

samples were taken, however, were not so closely equivalent as those from layer D. From toward the base of the layer, Sabels finds that cool, humid conditions were reflected by trace elements in the sample from 4.85 m. depth. This sample may be about 35,000+ years B.P. Leroi-Gourhan finds a definitely dry period reflected in her pollen sample from the 4.0-m. depth. This sample may be in the neighborhood of 33,000 B.P. There is an evident alternation from cool to warm climate. Furthermore, another change to a cool climate occurred near the top of this layer, according to Leroi-Gourhan. This appears to be close to the encroaching cold and wet conditions of the Late Würm in Europe, a long, cold stage about 15,000 years duration (R. F. Flint and F. Brandtner, 1961, p. 322).

Mention has been made that there is a cultural and chronological gap between layers B and C, which may be due to the abandonment of the cave during the Late Würm equivalent. In so far as we know from the evidence at present, the cave was not occupied following the Baradostian until about 12,000 years ago. At this time the climate, according to the findings of both Sabels and Leroi-Gourhan, was cooler and moister than it is today. Leroi-Gourhan indicates that this was the situation at both the cave and in the corresponding horizon at the village site. From about 10,600 years ago, both researchers show that the climate seems to have become warmer and drier to the present prevailing conditions.

It is very evident that the altitudinal zonation or life zones for the Zagros Mountain region as suggested by Wright (in R. J. Braidwood, and B. Howe, 1960, pp. 89–90) is very convincingly demonstrated in the data above. The snow line in the mountains possibly descended to 4,000 feet during the last glaciation, depressing the tree line below the level of Shanidar Cave (altitude 2500 feet). The cave was presumably inhospitable during this interval. Mention has been made of date palm-pollen grains in the Shanidar Cave deposits. Today date palms occur far to the south of this area. The farthest north they are found is in the region of Anah in central Iraq, at about 33° N lat. (W. D. Fisher, 1950, p. 356). In order to support date palms in the region of Shanidar in the geologic past, the temperature was presumably at least comparable to the present temperatures found at Anah. There at present, the mean daily July maximum temperature is about 110° F., and the mean daily January minimum temperature is about 35° F. At present, in the region of Shanidar, the mean daily July maximum temperature is about 95° F., and the mean daily January minimum temperature is about 25° F. (D. W. Fisher, 1950, p. 44). Thus it would appear to indicate that, assuming there were no altering factors, the temperatures at Shanidar during the time of the date palms in the Baradostian and Mousterian horizons must have been close to the present habitat of date palms in order to permit their growth in the Shanidar environs. The rainfall of this region at present, up to about 40 inches annually, must have been rather less during this time.

Concerning the flora of the present in the Zagros Mountain area of this region, we note that the scrub or dwarf oak is the principal wild tree there (H. Field, 1952, pp. 13, 31; W. R. Hay, 1921, p. 106). W. R. Hay (1921, p. 33) says that he observed juniper trees in one spot only, and that conifers were

not to be seen. There is no mention of date palms. Other trees present in the area include the sumach, hawthorn, plane, ash, walnut, poplar, mulberry, and willow.

Conclusion

We seem to have the first reliable record of palaeoclimate in the Near East, combined with archaeological data and based upon carbon-14 dates, which may be checked against the better known European system. From our information at present, we can speak with some assurance concerning the changes in climate extending back to at least 50,000 years, and more if the dating can be extended or extrapolated accurately by one or another means. It is expected that there will be additional refinements upon the present set of data as researches progress on the study of the Shanidar soil samples.

References

BRAIDWOOD, R. J. & B. HOWE. 1960. Prehistoric Investigations in Iraqi Kurdistan. Studies in Ancient Oriental Civilization No. 31. Univ. Chicago. Chicago, Ill.

BUTZER, K. W. 1958. Quaternary Stratigraphy and Climate in the Near East. Bonner Geographische Abhandlungen 24. Bonn.

FIELD, H. 1952. The Anthropology of Iraq. Papers of the Peabody Museum of Archaeology and Ethnology, Harvard Univ. Press. **XLVI**(2 and 3). Cambridge, Mass.

FISHER, W. D. 1950. The Middle East. Methuen. London, England.

FLINT, R. F. & F. BRANDTNER. 1961. Climatic changes since the last Interglacial. Am. J. Science. **259:** 321–328.

GARROD, D. A. E. & D. M. A. BATE. 1937. The Stone Age of Mount Carmel. **I.** Oxford Univ. Press. Oxford, England.

HAY, W. R. 1921. Two Years in Kurdistan. Sidgwick & Jackson, Ltd. London, England.

HOWELL, F. C. 1959. Upper Pleistocene stratigraphy and early man in the Levant. Am. Phil. Soc. Proc. **103:** 1–65.

MCBURNEY, C. B. M. 1960. The Stone Age of Northern Africa. Penguin Books. London.

SOLECKI, R. S. 1952. A Paleolithic site in the Zagros Mountains of Northern Iraq. Report on a sounding at Shanidar cave. Pt. 1. Sumer. **8:** 127–161.

SOLECKI, R. S. 1953*a*. A Paleolithic site in the Zagros Mountains of Northern Iraq. Report on a sounding at Shanidar cave. Part 2. Sumer. **9:** 60–93.

SOLECKI, R. S. 1953*b*. The Shanidar cave sounding, 1953 season, with notes concerning the discovery of the first Paleolithic skeleton in Iraq. Sumer. **9:** 229–239.

SOLECKI, R. S. 1955*a*. Shanidar cave, a Paleolithic site in Northern Iraq. Ann. Rept. Smithsonian Inst. for 1954. **:** 389–425.

SOLECKI, R. S. 1955*b*. Shanidar cave, a Paleolithic site in Northern Iraq. Sumer. **11:** 14–38.

SOLECKI, R. S. 1959. Early man in cave and village at Shanidar, Kurdistan, Iraq. Trans. N. Y. Acad. Sci. Ser. 2. **21**(8)**:** 712–717.

SOLECKI, R. S. 1960. Three adult Neanderthal skeletons from Shanidar cave, Northern Iraq. Ann. Rep. Smithsonian Inst. for 1959. **:** 603–635.

SOLECKI, R. S. & M. RUBIN. 1958. Dating of Zawi Chemi on early village site at Shanidar, Northern Iraq. Science. **127**(3312)**:** 1446.

STEKELIS, M. & G. HAAS. 1952. The Abu Usba Cave (Mount Carmel). Israel Exploration J. **2:** 15–47.

VAUFREY, R. 1938. Review: Garrod, D. A. E. & D. M. A. BATE. The Stone Age of Mount Carmel. Vol. 1. Excavations in the Wady el-Mughara. Anthropologie. **48:** 568–576.

VAUFREY, R. 1939. Paléolithique et Mésolithique Palestiniens. Revue Scientifique de la France et de l'Etranger VI–VII. **:** 390–406.

ZEUNER, F. E. 1958. Dating the Past. Methuen. London, England.

Printed in the United States of America

MONOGRAPHIC PUBLICATIONS

OF

THE NEW YORK ACADEMY OF SCIENCES

(Lyceum of Natural History, 1817–1876)

(1) The Annals (octavo series), established in 1823, contain the scientific contributions and reports of researches, together with the records of meetings of the Academy. The articles that comprise each volume are printed separately, each in its own cover, and are distributed immediately upon publication. The price of the separate articles depends upon their length and the number of illustrations, and may be ascertained upon application to the Executive Director of the Academy.

Current numbers of the Annals are sent free to all Members of the Academy desiring them.

(2) The Special Publications, established in 1939, are issued at irregular intervals as cloth-bound volumes. The price of each volume will be advertised at time of issue.

(3) The Memoirs (quarto series), established in 1895, are issued at irregular intervals. It is intended that each volume shall be devoted to monographs relating to some particular department of science. Volume I, Part 1 is devoted to Astronomical Memoirs, Volume II to Zoological Memoirs. No more parts of the Memoirs have been published to date. The price is one dollar per part.

(4) The Scientific Survey of Porto Rico and the Virgin Islands (octavo series), established in 1919, gives the detailed reports of the anthropological, botanical, geological, paleontological, zoological, and meteorological surveys of these islands.

Subscriptions and inquiries concerning current and back numbers of any of the publications of the Academy should be addressed to

EXECUTIVE DIRECTOR
The New York Academy of Sciences
2 East Sixty-third Street
New York 21, N. Y.